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ON THE CRYSTAL STRUCTURE OF THE CLAY MINERALS: DICKITE, HAL- LOYSITE AND HYDRATED HALLOYSITE

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DICKITE, $\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O}$.

Four distinct clay minerals, kaolinite, nacrite, dickite,¹ and halloysite,² are formed by superposition of layers having the ideal composition, $[(\text{OH})_4\text{Si}_2\text{Al}_2\text{O}_5]_n$. The structure within the layer was suggested by Pauling,³ and Gruner⁴ using powder photographs worked out more detailed structures for the first three minerals. His result for dickite has been questioned by Ksanda and Barth⁵ who obtained excellent data from single crystals. Mr. Ksanda of the Geophysical Laboratory, Carnegie Institution of Washington, has kindly furnished unpublished x-ray diffraction data on dickite for this further study.

Gruner, and Ksanda and Barth are in essential agreement on the dimensions of the unit of structure. Ksanda and Barth, however, consider the mineral to be monoclinic holohedral since it failed to show a piezoelectric effect, the suggested space group being C_{2n}^6-C2/c . This possibility is eliminated since single crystals show a strong pyroelectric effect when tested by the Martin method.⁶ Since the absent reflections require a glide reflection plane the space group must be C_s^4-Cc as found by Gruner.

That the structure is really formed by polar superposition of $[(\text{OH})_4\text{Al}_2\text{Si}_2\text{O}_5]_n$ sheets is shown by the intensities of $(00l)$ reflections. Barth and Ksanda noted up to the 20th order from this plane and showed that calculated and observed intensities were in moderate agreement. The following parameter values lead to excellent agreement as shown by Table 1; all atoms in general positions with c parameters for Si_1 , $\text{Si}_2 = .00$; O_3 , O_4 , $\text{O}_5 = .037$; O_1 , O_2 , $(\text{OH})_1 = -.110$; $(\text{OH})_2$, $(\text{OH})_3$, $(\text{OH})_4 = -.258$ and Al_1 , $\text{Al}_2 = -.191$. It follows from these parameter values and the close approach of the a_0 and b_0 dimensions to the ideal values that the structure of the layer must be close to that suggested by Pauling.

TABLE 1. OBSERVED AND CALCULATED INTENSITIES FOR (00 l) OF DICKITE, CuK α RADIATION

Order of reflection	Calc. Int.	Order of intensities		Estimated intensities
		observed	calculated	
2	53.0	2	2	9
4	77.0	1	1	10
6	13.7	3	3	5
8	13.0	4	4	5
10	6.9	6-7	6	3
12	4.1	6-7	7	3
14	2.2	8	8	2
16	10.5	5	5	4
18	0.4	9	9	1

There are four possible ways of superimposing the $[(\text{OH})_4\text{Si}_2\text{Al}_2\text{O}_5]_n$ layers, neglecting translations along the y axis, and these are schematically represented by figure 1, where the arrows represent reflection planes of the layers. The layers moreover can be shifted either by $+x$ or $-x$ along the a axis, the value but not the sign of the shift being fixed by the crystallographic angle β . Gruner's suggested dickite and kaolinite

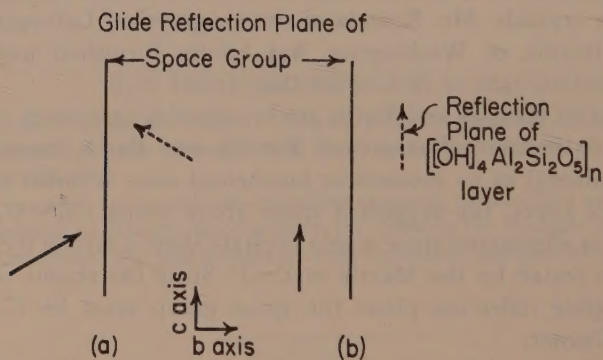


FIG. 1

structures are of type 1, and for convenience the first is designated by Aa ; the second by Bb . Shift of the aluminum atoms alone by $\frac{1}{3}b_0$ within the layers gives structures Aa' and Bb' of type 2.

Arrangements Bb and Bb' require $(400) > (200)$ and thus are eliminated for dickite. Both Aa and Aa' correctly account for the intensities of the $(h00)$ and $(20l)$ reflections as Barth and Ksanda have shown. This is also true for the $(40l)$ reflections save that $(40\bar{6})$ absent has too high a

calculated intensity, which, however is quite sensitive to choice of parameters. A most striking feature of the data from dickite is the absence of reflections from $(06l)$, with l odd and from $(h3l)$ with l even. This and the high intensity of (060) requires the layers to be shifted by $(2n+1)/6$ along the b axis with respect to each other. Moreover, it is to be noted that intensities of reflections for $(h3l)$ with l odd, calculated for any value of n (Ksanda and Barth's Table 8) are in approximate agreement with the observed values.

Intensities of $(0kl)$ reflections unfortunately cannot be explained by any choice of n . This can be seen by consideration of (020) , (021) , and (022) , Table 2, which have approximately the same intensities and values of $\sin \theta$. One is thus left in the position of having eliminated all possible structures strictly compatible with space group requirements.

TABLE 2. CALCULATED VALUES OF $F^2/16 F^2_{A1}$ FOR SOME PLANES OF DICKITE

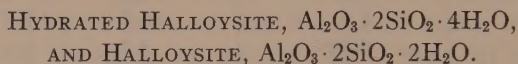
Plane	(020)	(021)	(022)
Intensity	ms	ms	ms
$Aa, n=1$	1.5	0.7	3.5
2	0.04	3.2	3.3
3	2.2	2.0	0.4
$Aa', n=1$	7.2	0.0	0.9
2	1.5	3.5	0.2
3	2.0	3.0	0.3
$S, n=1$	2.9	0.0	2.6
2	0.7	2.0	0.6
3	0.9	2.0	0.8

The only escape seems to be that the actual structure is one having an average arrangement required by space group Cc . In any case the averaged structure would be one based upon arrangements Aa and Aa' with shifts of one layer over the other in the b direction of $(2n+1)/6$. The radiation scattered from different parts of the crystal might or might not be coherent, thus the possibility must be borne in mind of summing amplitudes or their squares.

A logical structure might be one in which the aluminum atoms were equally distributed in the positions of Aa and Aa' . Intensities calculated assuming coherence of scattered radiation are shown under S of Table 2, and they are not satisfactory. Statistical structures with all values of n equally probable would require (hkl) with k odd to be absent if the scattered radiation was coherent. Structures Aa or Aa' with all possible shifts equally probable but without coherent scattering

from parts of the crystal of different structure would explain the (*hkl*) intensities. Intensities of (020), (021), and (022) are explained by *Aa* on this basis but not by *Aa'*, the latter making (022) far too weak as can be seen by summing *Aa'*, *n* = 1, 2, and 3 of Table 2. However, calculations for (023) and (027) eliminated *Aa* since the calculated intensities are too great.

Many combinations were tried with assumed coherence of the scattered radiation but none of these was satisfactory. It would seem, therefore, that the presumptive evidence is in favor of structure *Aa*, which was selected by Gruner, but *Aa'* is not eliminated. In the former structure, as Gruner has pointed out, the hydroxyl groups in the bottom of one layer would be equidistant from two oxygen atoms on top of another layer irrespective of the value of *n*. In any case a statistical structure would not be unexpected since forces between the layers are quite small. This is rather similar to the observation that CdBr_2 and NiBr_2 can form structures^{7,8} in which the layers are superimposed in a random manner while preserving a close packing of the large bromine ions.



The relationship of halloysite to the other clay minerals has been discussed by Mehmel.² Hofmann, Endell and Wilm⁹ recognized that a mineral of the ideal composition $\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 4\text{H}_2\text{O}$ could be obtained but that it readily dehydrates even at 50°C. Mehmel called the dehydrated product "metahalloysite" and indicated that its structure was formed by superposition of $[(\text{OH})_2\text{Si}_2\text{Al}_2\text{O}_5]_n$ layers as are the other kaolin minerals. The great ease with which this dehydration takes place has not generally been recognized and as a consequence there is some confusion in the literature, but as a rule Mehmel's "metahalloysite" is the usual halloysite of others.

In attempting to verify Mehmel's experimental results a number of x-ray powder diffraction photographs were made of "halloysite" samples kindly placed at my disposal by Dr. Clarence S. Ross, of the United States Geological Survey. The specimens, many of which had been in the laboratory for more than a year, had predominantly the diffraction pattern of kaolinite. Samples 5, 6, 10, 11, 14, 16, and 21 described by Ross and Kerr¹⁰ as halloysite according to microscopic examination, proved to be kaolinite, while samples 2, 8, and 9 were halloysite. Inherent difficulty in differentiating the two by microscopic methods was recognized by Ross and Kerr. A sample of "halloysite" from Anamosa Iowa, supplied by Dr. Ross, while predominantly halloysite, contained some of the hydrated material described by Hofmann, Endell and Wilm.

Professor Hofmann of the Chemical Institute of the University of Rostock most kindly sent me a sample of halloysite from Djebel Debar. Its diffraction pattern agreed entirely with that of hydrated halloysite as described by Mehmél² and shown by Hofmann, Endell and Wilm.⁹ Moreover, heating to 50° overnight, or exposure for a week to the laboratory air, was sufficient to change the diffraction pattern to that of halloysite. Neither the samples of halloysite from Dr. Ross' collection nor the dried material from Djebel Debar formed hydrated halloysite upon three months' standing in water at room temperature.

TABLE 3. ELECTRON DIFFRACTION DATA FROM HALLOYSITE

Spacing in Å units	Intensity ¹	Probable indices
4.35	vs	(110)(020)
2.56	s	(200)(130)
	A region of broad scattering	
1.654	m	(310)(240)(150)
1.483	ms	(060)(330)
1.298	m	(260)
1.226	m	(170)(350)(420)
1.105	w	(030)(440)
1.023	vw	(230)(370)
0.973	mw	(190)
0.956	mw	(460)(530)
0.872	w	(550)
0.842	vw	(390)(600)(2.10.0)
0.801	vw	(1.11.0)

¹ Abbreviations: v, very; s, strong; m, medium; w, weak.

Electron diffraction data from halloysite are summarized in Table 3. These results were obtained after the manner described in similar work on kaolinite and thus only reflections from (*hk*0) appear. Patterns of halloysite and kaolinite while closely similar have some distinct differences, particularly at the larger angles of scattering. Shishacow¹² has recently reopened the discussion as to whether or not such data are obtained from single layers, which he prefers to consider (Si₂O₅)_n rather than [(OH)₄Al₂Si₂O₅]_n. He, however, apparently was not familiar with the experiments on anauxite¹³ and mica¹⁴ which demonstrated why only (*hk*0) reflections appear from superimposed layers. Since Mehmél observed (*hkl*) reflections, by *x*-ray diffraction, and these were verified, the question of how the [(OH)₄Al₂Si₂O₅]_n layers are superimposed in halloysite is still open.

Electron diffraction photographs were taken using the diffraction pat-

tern of gold for reference and spacings are accurate to about 0.5 per cent on this basis. Shishacow¹² contends that use of gold for reference can lead to errors as great as 1 per cent in the wave length, but the point is not accepted.

Mehmel suggested that the structure of hydrated halloysite, $\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 4\text{H}_2\text{O}$ consists of $\text{Al}_2(\text{OH})_6$ and $\text{Si}_2\text{O}_5(\text{OH})_2$ layers superimposed with the symmetry of the space group C_s^3 — Cm . Dehydration to halloysite was assumed to take place by interaction of hydroxyl groups in one layer with those of another. This appears unlikely since such reactions generally require a large activation energy and thus would not be expected to take place readily at such low temperatures. The reaction would be analogous to the dehydration of gibbsite which takes place at a quite high temperature.

An alternative structure to that suggested by Mehmel would be one in which neutral $[(\text{OH})_4\text{Si}_2\text{Al}_2\text{O}_5]_n$ layers similar to those of the other clay minerals are interleaved with layers of $[2\text{H}_2\text{O}]_n$ molecules. Such a structure would be similar to that suggested for the vermiculites by Gruner.¹⁵ A choice between the two possibilities can be made from con-

TABLE 4. CALCULATED INTENSITIES FOR SOME ORDERS OF REFLECTION FROM (00 l) OF HYDRATED HALLOYSITE

Order	$\frac{\sin \theta}{\lambda}$	C	$F^2/4 \times C^1$		Observed intensity
		$\frac{1 + \cos^2 2\theta}{\sin^2 \theta \cos \theta} \times 10^{-3}$	Mehmel	Hendricks	
n	.0495	.350	110	720	vs
$2n$.099	.085	172	7	a
$3n$.1485	.040	18	73	
$4n$.198	.019	31	26	

Values of z parameters

Mehmel		Hendricks	
4Al	at	4Al	at
6(OH)	.118	2(OH)+4 O	.11
6(OH)	-.118	6(OH)	-.11
4Si	-.445	4Si	.27
4OH	.40	6O	.33
6O	-.375	4H ₂ O	.61 ²

¹ Scattering factors used are those listed in "The International Tables for Determination of Crystal Structures," Chemical Catalogue Company, New York, 1935.

² This value of the H_2O z parameter places the water molecules midway between (OH) groups and oxygen atoms along the axis. Its exact value does not greatly affect the calculated intensities, particularly for n and $2n$ reflections.

sideration of the intensities of x-ray reflections from (00 l). Relative calculated intensities of reflections according to the two postulates are shown in Table 4; parameter values also being listed.

An intense reflection from a plane with a spacing of 10.1 Å is observed on powder photographs of hydrated halloysite. The large value of this spacing, its change upon dehydration, and its orientation in samples of hydrated halloysite formed by settling from water, all suggest that it is an order of (00 l). Mehemel called it the first order reflection consistent with the requirements of the space group C_s^3 — Cm . The second order reflection is absent. Now Mehmel's structure requires a weak first order reflection and a stronger second order one; his calculated intensity for (00 l) is apparently in error. The alternative structure correctly explains these intensities. Data are hardly sufficient to permit further analyses of the structures of halloysite or hydrated halloysite.

SUMMARY

Dickite is shown to have an atomic arrangement based on space group C_s^4 — Cc . Independent $[(OH)_4Si_2Al_2O_5]_n$ layers are superimposed in some random manner in which a shift of $(2n+1/6)b_0$ of one layer with respect to another is preserved.

Hydrated halloysite, $Al_2O_3 \cdot 2SiO_2 \cdot 4H_2O$, probably has a structure in which neutral $[(OH)_4Al_2Si_2O_5]_n$ layers are interleaved with $[(H_2O)_2]_n$ layers.

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AMPHIBOLIZATION OF SILLS AND DIKES IN THE LIBBY QUADRANGLE, MONTANA*

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ABSTRACT

Numerous mafic sills, a few dikes, and one stock intrusive in the Belt rocks in the Libby quadrangle, Montana, show widespread amphibolization similar to that observed in the pre-Cambrian Purcell sills of British Columbia and Northern Idaho. The Montana sills are the same age as the Purcell intrusives, whereas the dikes and the stock are regarded as late Mesozoic. The amphibolization of both groups of intrusives is believed to have been caused by hydrothermal solutions which were derived from and followed the quartz monzonite, granodiorite, and similar intrusives that invaded the Belt rocks of northern Idaho and northwestern Montana, probably in the late Mesozoic. The ore deposits in these areas also are genetically related to the granodiorite and quartz monzonite.

INTRODUCTION

A series of mafic sills, commonly called the Purcell sills, intrusive into the Belt sedimentary rocks of northern Idaho and southern British Columbia has been described by different workers¹ and assigned to the pre-Cambrian.

* Published with the permission of the Director, United States Geological Survey.

¹ Daly, R. A., *Geology of the North American Cordillera at the 49th parallel: Mem. 38, Part I, Canada Geol. Survey, Dept. of Mines*, pp. 209-255, 1912.

Schofield, S. J., *The origin of granite (micropegmatite) in the Purcell sills: Canada Geol. Survey, Mus. Bull. 2*, pp. 1-32, 1914.

Schofield, S. J., *Geology of the Cranbrook map-area, British Columbia: Mem. 76, Canada Geol. Survey, Dept. of Mines*, pp. 56-70, 1915.

Calkins, F. C., *Geological reconnaissance in northern Idaho and northwestern Montana: U. S. Geol. Survey, Bull. 384*, pp. 48-50, 1909.

In the course of the mapping of the Libby quadrangle, Montana, (Fig. 1) for the United States Geological Survey, many similar sills in

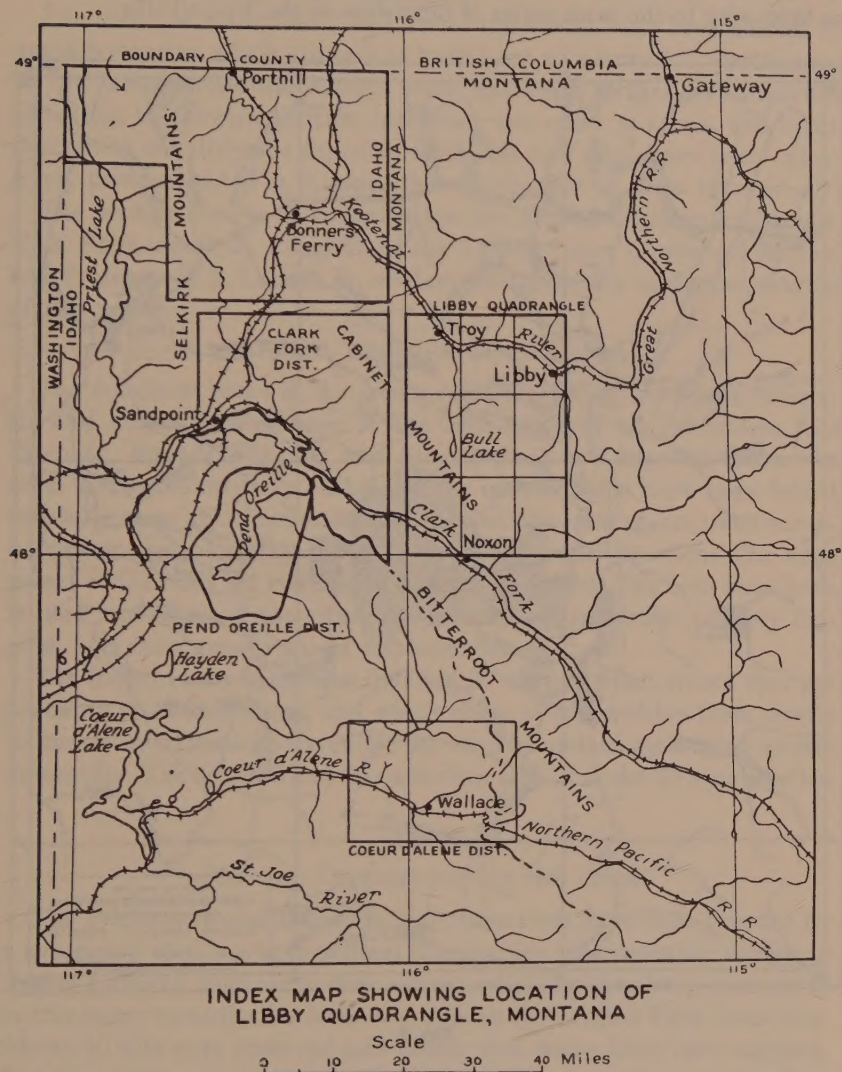


FIG. 1

Kirkham, V. R. D., and Ellis, E. W., Geology and ore deposits of Boundary County Idaho: *Ida. Bur. of Mines and Geology, Bull. 10*, pp. 36-38, 1926.

Anderson, A. L., Geology and ore deposits of the Clark Fork district, Idaho: *Ida. Bur. of Mines and Geology, Bull. 12*, pp. 24-25, 1930.

the Belt rocks of that quadrangle and adjacent areas north and south of the Libby quadrangle were mapped and examined. They are regarded as belonging to the same series of intrusives as the Purcell sills.

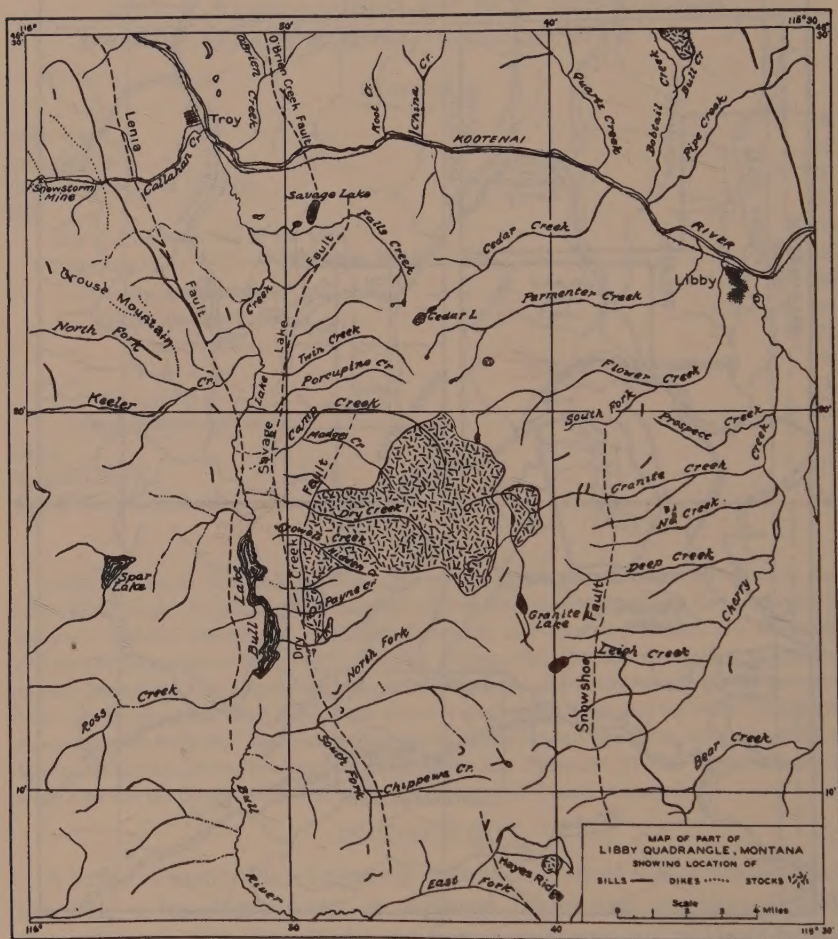


FIG. 2

A number of mafic dikes and a few felsic dikes accompanied, or succeeded, the quartz monzonite and granodiorite stocks² which invaded the

² Gibson, Russell, Campbell, Ian, and Jenks, W. F., Quartz monzonite and related rocks of the Libby Quadrangle, Montana, and the effects on them of deuteritic processes (in preparation).

Belt rocks probably in the late Mesozoic. A few of the sills may possibly belong to this later period of intrusion.

The sills, the mafic dikes, and one of the stocks have been altered by amphibolization which appears to have been widespread. Superimposed on the amphibolization locally there has been further hydrothermal alteration to chlorite, sericite, carbonate and other minerals, and introduction of metalliferous deposits. Both sills and dikes have been extensively prospected.³ (Fig. 2 shows the distribution of these intrusives.^{3a})

The writers desire to acknowledge the helpful criticism of Messrs. C. S. Ross and G. F. Loughlin of the United States Geological Survey, and Professor E. S. Larsen, Jr. of Harvard University in the preparation of this paper.

GENERAL GEOLOGY OF THE AREA

The dominant rocks of the area are shale, argillite, sandstone, quartzite and dolomitic limestone of the Belt series. Four formations, the Prichard, Ravalli, Wallace, and Striped Peak are exposed within the area. These sedimentary rocks and the included sills have been folded into large open anticlines and synclines which in most places trend north-northwest and plunge at a low angle in the same direction. The axial planes of the folds are commonly inclined eastward. The beds have been faulted and many of the crosscutting faults are persistent, steeply dipping, and show great vertical displacement.

The folded rocks have been invaded by several small stocks that are probably of Mesozoic age, and which range in composition from diorite to quartz monzonite and syenite; the most abundant rock being quartz monzonite. The largest stock has a surface area of about 20 square miles.

SILLS

SIZE AND DISTRIBUTION OF THE SILLS

The Prichard, Ravalli, and Wallace formations have been invaded by a succession of sills whose original composition was probably similar to that of a diorite. Because of their altered condition they have been called in this paper metadiorite. None was seen in the Striped Peak formation. About 40 sills were observed and nearly that many have been mapped. In some instances the same sill may possibly have been mapped on both limbs of a fold and regarded as two sills. Most of them are thin and

³ Gibson, Russell, Geology and ore deposits of the Libby quadrangle, Montana: *U. S. Geol. Survey, report in preparation*.

^{3a} The Dry Creek fault shown on the map is the Bull Lake fault of Calkins, *U. S. Geol. Survey, Bull. 384*, pp. 67, 68, 1909.

discontinuous and cannot be traced far; and, as individual sills show very different degrees of alteration, correlation is difficult.

The longest and most conspicuous sills are in the northwestern part of the quadrangle between Keeler Creek and Preacher Mountain west of Troy; in the northeastern part on Mount Sheldon; and in the southeastern part east and west of the Snowshoe fault. Small sills were observed in many different places within the quadrangle. They decrease in number and thickness toward the south.

The sills range in thickness from 10 inches to 800 feet and in length from a few feet to 10 miles. Most of them are about 100 feet or less in thickness and only five have been traced for a distance greater than two miles. They weather easily and do not commonly form conspicuous features of the topography, and hence, their dimensions may commonly be greater than those recorded.

About 14 per cent of the total number of sills are in the Prichard, 16 per cent in the Ravalli, and 70 per cent in the Wallace formation, but the greatest thickness of sill material, roughly 1200 feet, is in the Prichard.

PETROGRAPHY

Megascopic features

The most abundant type is a holocrystalline, medium-grained, dark-green metadiorite composed chiefly of hornblende and tabular plagioclase, with smaller amounts of biotite and accessory minerals. Quartz rarely exceeds 15 per cent and in a very few sills it is lacking. The average grain size of hornblende, the dominant mineral, is 1.4 millimeters; of the plagioclase 0.6 millimeter. These interlock in random orientation in a tight fabric.

A second type, which is of minor importance quantitatively, is a porphyritic rock in which phenocrysts of plagioclase, quartz, or mafic minerals are conspicuous in an abundant groundmass similar in grain-size and composition to the finer-grained facies of the metadiorite. These types grade into each other.

The metadiorite sills differ among themselves and in different parts of the same sill in grain size, in proportions of the essential minerals, and in degree of alteration. No stratification was observed according to mineral density nor were any pegmatite streaks seen.

Hornblende is completely, or almost completely, replaced in over half the sills. In some of these little or no plagioclase remains, and the rock is composed chiefly of minerals mentioned below which are the result of further hydrothermal alteration. This is true especially near prospects and mines where metalliferous veins have been opened up in the sills.

Modes of several sills are given in Table 1.

TABLE 1

	1	2	3	4	5	6	7	8	9	10
Hornblende	59.6	64.6	53.4	tr	18.3	18.4	11.4	—	—	—
Plagioclase	17.1	12.8	23.4	2.3	19.3	30.1	18.7	51.8	42.5	57.6
Quartz	14.0	5.8	5.8	14.3	2.3	2.1	3.4	0.4	9.8	—
Biotite	1.0	1.5	—	12.9	11.7	16.6	8.8	1.0	18.1	—
Chlorite	1.0	4.4	3.9	28.4	19.2	8.6	22.3	28.9	19.6	35.0
Carbonate	0.3	0.7	2.3	17.5	2.7	4.6	8.5	—	2.8	—
Sericite	tr	3.3	1.9	17.5	tr	3.2	4.8	tr	—	tr
Epidote and clinozoisite	3.6	3.6	6.0	tr	17.8	9.9	13.3	tr	tr	tr
Accessory	3.4	3.3	3.3	7.1	8.7	6.5	8.8	17.9	7.2	7.4
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

1. Long sill between the Lenia fault and the Snowstorm mine.
2. Less altered part of sill at Liberty mine, south of Callahan Creek.
3. Sill at Paul ranch, south of the Liberty mine sill.
4. Sill at Silver King mine on North Fork of Keeler Creek.
5. Sill on Burnt Mountain on north side of Keeler Creek.
6. Sill at Martin's prospect, northwest of Hayes Ridge.
7. Long sill from Bear to Leigh Creeks.
8. Sill on Mt. Sheldon between Kootenai River and Pipe Creek.
9. Sill on North Fork of Bull River.
10. Sill on Horse Creek, north of Granite Creek.

Microscopic features

Several textures and great differences in grain size are seen under the microscope. The most common texture is controlled by long interlocking euhedrons of hornblende. Granular texture is seen only in the thoroughly altered specimens. Where foliation is present it is formed largely by reorientation of biotite and by chlorite. A porphyritic texture with a matrix that is coarse for a true porphyry is found in a few sills.

Next to hornblende and plagioclase the most abundant minerals are chlorite, carbonate, sericite, quartz and biotite, which are chiefly alteration minerals. Epidote, and clinozoisite are erratic in amount and distribution. Orthoclase, sphene, rutile, leucosene, apatite, magnetite, ilmenite, tourmaline, and zircon are the accessory minerals.

Hornblende. Hornblende ranges in size from 0.02 to 18 millimeters and averages about 1.4 millimeters. It is strongly pleochroic, and forms euhedrons and subhedrons. In places the fabric of the rock is controlled by elongate interlocking hornblende grains which crystallized later than the smaller plagioclase grains and, consequently, enclose, penetrate, or wrap around the plagioclase (Fig. 3). According to Harker⁴ this texture is markedly developed in rocks in which the hornblende is in great part

⁴ Harker, Alfred, *Petrology for students*, pp. 67-68, Cambridge, 1923.

derived from augite. Augite was observed in only one sill and here it was partly replaced by hornblende.



FIG. 3. Sketch of part of thin section showing crosscutting nature of hornblende. Magnification 30X.

Optical properties of hornblende from several sills are given in the accompanying Table 2. Hornblende with lower indices of refraction is found in intrusives in association with plagioclase of more calcic composition, whereas hornblende of higher indices is associated with more sodic plagioclase. Rice⁵ found this same association in the Purcell sills of British Columbia. Optical properties of two hornblendes are quoted from Rice's paper for comparison. His analyses of these two hornblendes show that Type "A," which is associated with the more sodic plagioclase, has slightly more potash and soda than Type "B"; whereas Type "B," which is associated with more calcic plagioclase, has slightly more lime than Type "A."

TABLE 2

Intrusive	α	β	γ	$\gamma - \alpha$	$Z/\angle c$	$2V$	sign
1.	1.637	1.652	1.661	.024	18°	75°	Neg.
2.	1.638	1.654	1.658	.020	18°	53°	Neg.
3.	1.658	1.672	1.681	.023	19°	77°	Neg.
4.	1.669	1.681	1.691	.022	16°	85°	Neg.
5.	1.666	1.678	1.688	.022	18°	84°	Neg.

⁵ Rice, H. M. A., Amphibole from the Purcell sills, British Columbia: *Am. Mineral.*, vol. 20, pp. 307-309, 1935.

Pleochroism		
X	Y	Z
1. Pale olive green	olive green	bluish green
2. Pale yellow	olive green	greenish blue
3. Pale yellowish blue	dark yellowish green	dark greenish blue
4. Pale yellowish brown	pale olive green	deep bluish green
5. Pale yellow	olive green	greenish blue

1. Dike at Togo prospect on Callahan Creek.
2. Purcell sill in British Columbia. Rice's type "B" hornblende.^a
3. Large sill between Snowstorm Mine and Lenia Fault.
4. Sill at Martin's prospect northwest of Hayes Ridge.
5. Purcell sill in British Columbia. Rice's type "A" hornblende.^a

^a Rice, H. M. A., *op. cit.*

Hornblende shows all degrees of alteration and is the first mineral to be attacked locally by hydrothermal solutions which deposited the ores. In some specimens, however, it is fresh and constitutes 65 per cent of the rock. It is replaced chiefly by chlorite and to a less extent by carbonate, epidote, biotite, magnetite and albite.

Plagioclase. Most of the feldspar of the sills is plagioclase, ranging in composition from albite to calcic andesine, the most abundant being oligoclase or andesine. In some specimens there are two kinds of plagioclase, but zoning is of minor importance. Oligoclase and andesine have been replaced very extensively by more sodic feldspar, albite or albite-oligoclase, accompanied by epidote and clinozoisite. The average grain size is 0.6 millimeter, but the common range is from about 0.1 to 2 millimeters. In exceptional cases the grains reach a length of 30 millimeters. Alteration of plagioclase to other minerals is not so widespread, so conspicuous, or so early as alteration of hornblende, and a rock in which the hornblende has been completely or almost completely replaced may contain 15 to 65 per cent albite and, less commonly, andesine. In the neighborhood of prospects, and to a minor extent elsewhere, the plagioclase is replaced by sericite, clinozoisite, epidote, and carbonate.

Quartz. Quartz is erratic in distribution and grain size but is commonly fine grained. Much of it has clearly been introduced during hydrothermal alteration of the sills containing metalliferous veins, but even in those sills which are still chiefly hornblende and plagioclase, and contain no metalliferous deposits, quartz commonly makes up 5 to 15 per cent. Anderson⁶ also mentions the usual amounts of quartz present in the Purcell sills of northern Idaho and southern British Columbia.

⁶ Anderson, A. L., Geology and ore deposits of the Clark Fork district: *Ida. Bureau of Mines and Geol., Bull.* 12, pp. 26-27, 1930.

CONTACT METAMORPHISM

Contact metamorphism near the sills is commonly inconspicuous. The effects of recrystallization are seen everywhere in the Belt rocks and it is not always possible to determine how much metamorphism is attributable to the sills and how much is regional. It seems clear that, near some sills, small amounts of biotite (later altered to chlorite), feldspar, quartz, carbonate, pyrite, pyrrhotite, magnetite, tourmaline and sphene have been disseminated in the enclosing argillaceous sandstone. So much biotite has been introduced into the sandstone a few inches from a sill contact, in a few places, that it is difficult to determine the exact position of the contact. Near any metalliferous deposits in a sill, metamorphism is a little more conspicuous in the enclosing sedimentary rocks. Here, sulphides and carbonate, especially, are more abundant in disseminated grains.

DIKES

The dikes have been described elsewhere⁷ and will be mentioned here only briefly.

Most of the large or persistent dikes in the Libby quadrangle are metadiorite and are in the northwestern part, north and south of Callahan Creek and west of the Lenia fault. About 10 dikes of this type were observed, two of which were more than 140 feet in thickness. The largest mine in the quadrangle, the Snowstorm, and many of the prospects in the Grouse Mountain area are in metadiorite dikes.

A few thin dikes which range in composition from granite to quartz diorite cut the quartz monzonite stocks and the sediments near the border of the stocks. All the dikes are distinctly crosscutting and are later than the folding.

The typical metadiorite dike is made up chiefly of hornblende and andesine. The plagioclase is less commonly sodic labradorite or oligoclase. From one to 10 per cent of quartz is present and the accessory minerals are apatite, zircon, tourmaline, sphene, magnetite and ilmenite. Where hornblende is fresh, little or no biotite is present. One of the dikes contains from 25 to 40 per cent of augite, but it is all partly altered to hornblende. In short, the metadiorite dikes are now roughly similar to the sills in composition and texture, though the range in grain size in the dikes is not so great as in the sills.

Many of the metadiorite dikes contain metalliferous deposits, and they are commonly more thoroughly altered than the petrographically similar sills, consequently some are now wholly or partly composed of chlorite, sericite, carbonate, quartz, and biotite, with smaller amounts

⁷ Gibson, Russell, Campbell, Ian, and Jenks, W. F., *op. cit.*

of clinozoisite, epidote and other minerals. The plagioclase in the sills is much albitized, but this has not been observed to so great an extent in the dikes.

AMPHIBOLIZATION

The mafic sills and dikes have been called metadiorite because it is believed that, like the sills in British Columbia and in the Clark Fork and Boundary County districts of northern Idaho, they have undergone widespread metamorphism. As mentioned above, the hornblende is clearly later than the plagioclase and may have been derived from a pyroxene, but in only one sill and in one dike did the hornblende grains show cores of augite.

In the Clark Fork district west of the Libby quadrangle Anderson⁸ found that hornblende, in similar sills believed to be pre-Cambrian, contains cores of hypersthene and augite, and suggests that hornblende in all these sills is a replacement of pyroxene. Kirkham and Ellis⁹ are of the same opinion concerning the hornblende in the pre-Cambrian sills in Boundary County, Idaho, northwest of the Libby quadrangle. The sills in these three nearly contiguous areas are very similar in composition, occur in the same formations of the Belt series, and are almost certainly identical in age. It is believed, therefore, that they have had very similar subsequent histories.

Schofield¹⁰ called attention to the great changes which have taken place in the pre-Cambrian Purcell sills of British Columbia, a part of the southern boundary of which is contiguous with Boundary County, Idaho. Here the augite and hypersthene of the original gabbro have been altered to hornblende so that the rock is now a hornblende gabbro. Schofield also mentions a quartz diorite phase of the sills in which the hornblende appears to be secondary. On the other hand, Daly,¹¹ who first described these sills, regards the rock as a primary hornblende gabbro, but he does not mention finding augite almost completely altered to hornblende, as do Schofield, Anderson, and Kirkham and Ellis.

Further evidence of widespread alteration of the sills in the Libby quadrangle is presented by the replacement of calcic plagioclase by albite, a process that is much more striking in the sills than in the dikes.

⁸ Anderson, A. L., Geology and ore deposits of the Clark Fork district, Idaho: *Ida. Bur. of Mines and Geol., Bull.* 12, p. 27, 1930.

⁹ Kirkham, V. R. D. and Ellis, E. W., Geology and ore deposits of Boundary County, Idaho: *Ida. Bur. of Mines and Geology, Bull.* 10, p. 37, 1926.

¹⁰ Schofield, S. J., Geology of Cranbrook map-area, British Columbia: *Canada Geol. Survey, Memoir* 76, pp. 56-70, 1915.

¹¹ Daly, R. A., Geology of the North American Cordillera at the 49th Parallel: *Canada Geol. Survey, Memoir* 38, pp. 222-223, 1912.

The metamorphism proceeded in stages of decreasing intensity. First the pyroxene was converted to amphibole. This was succeeded by albitization of plagioclase, in places accompanied by the formation of epidote and clinozoisite. Biotite begins to appear as hornblende diminishes in amount, though some biotite seems to be contemporaneous with hornblende. Later, both hornblende and biotite are replaced by chlorite and other minerals. Finally, plagioclase disappears and the rock is composed of sericite, clinozoisite, epidote, carbonate, chlorite and quartz. This complete alteration is conspicuous especially where metalliferous veins have been deposited in the intrusives.

As alteration proceeds, the texture of the rock changes. The hornblende-plagioclase rock is fine- to coarse-grained dioritic; the thoroughly altered rock is fine grained and shows little or no trace of its former texture. There are all gradations between these two textural types.

The writers believe that clues which account for the widespread amphibolization and albitization are found in the numerous prospects and mines in northern Idaho and northwestern Montana; and in the numerous, though in places, small intrusives of quartz monzonite and closely related rocks. These intrusives are less abundantly exposed in northwestern Montana than in northern Idaho, but the scattered small stocks in Montana are probably cupolas on larger intrusives not far below the present surface. If this be granted, it is not difficult to conceive of hydrothermal solutions from these intrusives which thoroughly permeated the rocks and brought about the amphibolization and albitization of the sills and dikes. To these same solutions may be ascribed other changes in the Belt rocks of the Libby quadrangle not discussed above, namely the albitization of certain limestones in the Wallace formation, the introduction of disseminated pyrite and pyrrhotite into many different beds of otherwise slightly altered Belt strata, and the filling of many small fissures with quartz veins. Some of these phenomena are seen in rocks remote from the outcrop of any known stock. Finally, these solutions deposited the ores seen in the prospects and mines of northwestern Montana and northern Idaho.

AGE OF THE SILLS AND DIKES

No definite evidence of the age of the sills and dikes is found within the Libby quadrangle. They are all in Belt rocks. No sills were seen in rocks younger than the Wallace, but dikes cut all of the Belt strata. Similar sills in the Clark Fork region¹² and in Boundary County, Idaho,¹³ are

¹² Anderson, A. L., *Geology and ore deposits of the Clark Fork district, Idaho: Ida. Bur. of Mines and Geology, Bull. 12*, pp. 25-26, 1930.

¹³ Kirkham, V. R. D., and Ellis, E. W., *Geology and Ore Deposits of Boundary County, Idaho: Ida. Bur. of Mines and Geology, Bull. 10*, pp. 36-38, 1926.

regarded as Algonkian and have been correlated with the Purcell sills in Canada described by Daly¹⁴ and Schofield.¹⁵ The sills are more numerous in these areas in northern Idaho than they are in the Libby quadrangle and farther south. Only one sill was seen in the Trout Creek quadrangle immediately south of the Libby quadrangle, and none is reported from the Coeur d'Alene district. Thus, in the areas under discussion in northern Idaho and northwestern Montana the sills appear to decrease in number and thickness from northwest to southeast.

The area is invaded by dikes of late Mesozoic (?) age, some of which are very similar in composition and appearance to the sills; and it is entirely possible that some of the sills are the same age as the dikes. However, the long sills, which appear to have been folded and faulted with the Belt rocks and to have taken part in all the orogenic movements which affected these rocks, are certainly to be correlated with similar pre-Cambrian sills in northern Idaho. Only one of the sills in the Libby quadrangle regarded as pre-Cambrian exhibited a crosscutting apophysis. This was a small dike and was traced for only a few feet. The greater part of the intrusive in question is clearly a sill.

The stocks are regarded as late Mesozoic¹⁶ because of their proximity to very similar stocks in Idaho, regarded as late Mesozoic. The dikes belong to the same period of intrusion as the stocks. The solutions responsible for the amphibolization and other related phenomena are believed to have soaked through the region shortly after the intrusion of the stocks and dikes. However, the sills are pre-Cambrian, and it is recognized that there may have been two periods of amphibolization, the earlier of which followed the intrusion of the sills but preceded the Mesozoic intrusives.

¹⁴ Daly, R. A., *Geology of the North American Cordillera at the 49th Parallel: Mem. 38, Part I, Canada Geol. Survey, Dept. of Mines*, pp. 207-255, 1912.

¹⁵ Schofield, S. J., *Geology of Cranbrook map-area, British Columbia, Mem. 76, Canada Geol. Survey, Dept. of Mines*, pp. 56-70, 1915.

¹⁶ Gibson, Russell, Campbell, Ian, and Jenks, W. F., *op. cit.*

UNGEMACHITE AND CLINO-UNGEMACHITE: NEW MINERALS FROM CHILE

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ABSTRACT

Ungemachite. System and lattice, hexagonal— R ; class, rhombohedral— $\bar{3}$; $a:c=1:2.2966$; $\alpha=62^\circ 51\frac{1}{2}'$; 32 forms. Habit, thick tabular $\{0003\}\{111\}$. Space group C_{6h}^2 — $R\bar{3}$. Hexagonal cell dimensions: $a_0=10.84\pm 0.02$ Å, $c_0=24.82\pm 0.05$ Å; $a_0:c_0=1:2.290$. Rhombohedral cell dimensions: $a_{rh}=10.37$ Å; $\alpha=62^\circ 59\frac{1}{2}'$. $V_{rh}=842$ cubic Å; $M_{rh}=1167$; contains $\text{Na}_8(\text{K}, \text{Fe}''')_4(\text{OH})_2(\text{SO}_4)_8\cdot 10\text{H}_2\text{O}$. Cleavage $\{0003\}\{111\}$, perfect and easy. Brittle; fracture irregular, with glassy luster. Hardness, $2\frac{1}{2}$. Density, 2.287 ± 0.003 . Colorless to yellowish; transparent. Uniaxial, negative. Indices of refraction (Na): $\omega=1.502\pm 0.002$, $\epsilon=1.449\pm 0.002$. Soluble in weak HCl. Analysis: Na_2O 21.61, K_2O 11.35, Fe_2O_3 7.69, SO_3 40.23, H_2O 16.69, N_2O_5 trace, insol. $2.07=99.64$. Composition, $\text{Na}_4(\text{K}, \text{Fe}''')_2(\text{OH})-(\text{SO}_4)_4\cdot 5\text{H}_2\text{O}$. Occurs with sideronatrite in massive altered iron sulphates at Chuquicamata, Chile. Named in memory of Henri Léon Ungemach [1880–1936] of Strasbourg.

Clino-ungemachite. System and lattice, monoclinic— F ; $a:b:c=1.6327:1:1.7308$; $\beta=110^\circ 40'$; 24 forms; the simple lattice is pseudo-rhombohedral, like that of ungemachite but definitely different. Visibly indistinguishable from ungemachite. Remaining properties not determinable for lack of material.

INTRODUCTION

In 1935 one of us (Bandy) spent four months in northern Chile collecting minerals for the United States National Museum and the Mineralogical Museum of Harvard University. The visit proved highly productive, yielding over a ton of material which consists mainly of specimens of the many natural salts of copper and iron for which the rainless region of Chile is famous. A number of well-known species are represented in the collection, also some rare species and a few that are certainly new. The collection was divided between the interested institutions and rapidly studied, first in Washington and later in Cambridge, in collaboration with the resident mineralogists.

The present paper concerns a pair of the new minerals which were studied in the Harvard Laboratory under the direction of Professor Charles Palache. In this work we were assisted in various ways by Doctor Berman, Doctor West and Mr. Gonyer, and especially by Professor Palache, who took a large part in the morphological study.

The two new species are much alike, consequently they were not at once distinguished; furthermore, their general properties appeared to agree with those given for an artificial salt identified with the mineral loweite— $\text{Na}_2\text{SO}_4\cdot\text{MgSO}_4\cdot 2\frac{1}{2}\text{H}_2\text{O}$. When the analysis of the more abundant substance gave a composition unlike that of any known salt it was

clear that we were dealing with a new mineral species. For this mineral, which is rhombohedral, we propose the name *ungemachite*, in memory of Henri Léon Ungemach [1880–1936] of Strasbourg, whose life-long devotion to crystal morphology has enriched the literature of descriptive mineralogy with many detailed studies, almost the last of which was an outstanding contribution (1935 A) on the natural sulphates of Chile. The rarer of the two new minerals appears to differ from *ungemachite* mainly in its symmetry, which is monoclinic; it is therefore named *clino-ungemachite*.¹

GEOMETRICAL RELATIONS IN THE HEXAGONAL SYSTEM

Goldschmidt's treatment of the hexagonal system, involving the alternative settings G_1 and G_2 , has given rise to ambiguities which have led to wide-spread difficulty and confusion.² In working out the morphology of *ungemachite* it was found that these ambiguities disappeared if only the G_1 setting is used and certain simple changes are made in the orientation of the polar axes and prime meridian. The resulting presentation is then completely consistent and in entire agreement with that used in the standard works of reference.

The hexagonal system, in the wide sense, comprises all crystals with a hexagonal or rhombohedral lattice. The rhombohedral lattice is a "centered" hexagonal lattice obtained by systematically adding two lattice points on a long body-diagonal of each hexagonal cell, as shown in the plan (Fig. 1) and the inclined view (Fig. 3). In accordance with accepted convention the hexagonal (Bravais) axes $A_1 A_2 A_3 C$ and the rhombohedral (Miller) axes $M_1 M_2 M_3$ take the directions shown. If we denote the simple (primitive) hexagonal lattice as hexagonal— P and the centered (rhombohedral) hexagonal lattice as hexagonal— R , and add the axial ratio $a:c$, which is common to both, the mode of the lattice and its geometrical form are completely defined.

The plan (Fig. 2) and inclined view (Fig. 4) show the corresponding reciprocal (polar) lattice in proper relative positions. The reciprocal rhombohedral lattice is obtained from the reciprocal hexagonal lattice by systematically omitting two points out of every three, as shown. The hexagonal polar axes $P_1 P_2 R$ (normals to the faces of the direct hexagonal lattice cell) and the rhombohedral polar axes $H K L$ (normals to the faces of the direct rhombohedral lattice cell) take the directions indicated; and again the lattice is correctly defined by a symbol indicating the lattice mode and the common polar axial ratio $p_0:r_0$.

¹ The word is purposely hyphenated to break the succession of vowels.

² These have been pointed out again by Parsons (1937).

Alternatively the rhombohedral lattice may be defined by the direct rhombohedral axial angle, $M_1:M_2=\alpha$ (Fig. 3), or by the reciprocal rhombohedral axial angle, $H:K=\lambda$ (Fig. 4). The four equivalent geometrical elements are related to the fundamental angle $\rho=(0001):(10\bar{1}1)$ by the equations:

$$\begin{aligned} \rho_0:r_0 &= \tan \rho : 1 \\ \sin \frac{1}{3}\lambda &= \frac{1}{3}\sqrt{3} \cdot \sin \rho \\ a:c &= 1:\frac{1}{3}\sqrt{3} \cdot \tan \rho \\ \tan \frac{1}{3}\alpha &= \sqrt{3} \cdot \cos \rho \end{aligned}$$

A rhombohedral lattice plane is correctly denoted by the Bravais symbol $(h \ k \ i \ l)$ where $h \ k \ l$ are the co-ordinates of the corresponding reciprocal lattice point on the polar axes $P_1 \ P_2 \ R$ and $i=h+k$. The hexagonal reciprocal lattice points which vanish in the rhombohedral reciprocal lattice are those whose co-ordinates are such that $h+i+l$ is not divisible by three; therefore it is clearly necessary to denote rhombohedral crystal planes by symbols that satisfy the law³ $h+i+l=3n$. Rhombohedral lattice planes may also be denoted by the Miller symbol $(h \ k \ l)$, where $h \ k \ l$ are the co-ordinates of the corresponding reciprocal lattice point on the polar axes $H \ K \ L$; and since the rhombohedral lattice is a primitive lattice the Miller indices are not subject to any extinction law. The indices $h \ k \ l$ (Bravais) and $h \ k \ l$ (Miller) are related by the transformation formula:⁴

$$\begin{aligned} \text{Bravais to Miller: } \frac{1}{3}\frac{1}{3}\frac{1}{3} / \frac{1}{3}\frac{1}{3}\frac{1}{3} / \frac{1}{3}\frac{1}{3}\frac{1}{3} \\ \text{Miller to Bravais: } 1\bar{1}0/01\bar{1}/111 \end{aligned}$$

When the hexagonal projection axes $P_1' \ P_2'$ are drawn parallel to the hexagonal polar axes $P_1 \ P_2$ on the plane of the gnomonic projection (the plane of the first layer of the reciprocal hexagonal lattice), then the co-ordinates $p_1 p_2$ of any face-pole in the double sextant enclosed by the positive Bravais axes $A_1 \ A_2$ are related to the corresponding Bravais and Miller indices as follows:

$$p_1 = h/l; \ p_2 = k/l \text{ (Bravais)}$$

$$p_1 = (h-k)/(h+k+l); \ p_2 = (k-l)/(h+k+l) \text{ (Miller)}$$

The Miller indices may be read directly from the gnomonic projection by a simple method given by Barker (1922, p. 80).

Finally, if we place the prime meridian midway between the positive axes $P_1 \ P_2$ then the face-poles with $\phi=0^\circ$ to $+60^\circ$ represent positive

³ This is Ungemach's (1935 B) convenient form of the law governing the four-index symbols of rhombohedral planes.

⁴ Written in the correct fractional form, as suggested by J. D. H. Donnay.

forms and face-poles with $\phi = 0^\circ$ to -60° represent negative forms, in accord with accepted convention. The positive and negative sextants divide into right and left half-sextants, permitting the maximum differentiation required in the lowest symmetry classes of the hexagonal system.

The treatment outlined is essentially the G_1 treatment of Goldschmidt with necessary changes in the positions of the axes $P_1 P_2 (P_1' P_2')$ and the prime meridian, arising from a clear recognition of the fact that the elements, symbols and angles must be interconsistent in the polar and linear presentations.

UNGEMACHITE

Morphology. The crystals of ungemachite occur singly or in groups of parallel individuals, associated with sideronatrite in friable massive altered iron sulphates from Chuquicamata, Chile. The single crystals are mostly less than 1 mm. in size, colorless or yellowish, distinctly rhombohedral in habit with a large base and numerous narrow faces developed on all the free sides. The crystals are easily brought into accurate adjustment by polar setting of the excellent basal reflection, the truncating faces falling mainly into three radial zones clearly revealing a three-fold, but not a six-fold symmetry axis. The symmetry class was determined from the relations of the third order terminal planes on both the upper and lower sides of single crystals. In several cases such planes were repeated only by the three-fold axis and an inversion center. This minimum symmetry was found sufficiently often to determine the class as rhombohedral— $\bar{3}$, in which the general form is the rhombohedron of the third order.

The gnomonic projection of the observed forms of ungemachite (Fig. 5) clearly shows that the lattice is rhombohedral. This may be seen in a qualitative way in the complete dissimilarity of adjacent vertical zones of first order forms; more definitely the rhombohedral character of the lattice is shown by the fact that the majority of the hexagonal face-symbols (Table 1) conform to the rhombohedral centering law: $h+i+l=3n$. The symbols of only four out of the thirty-two forms, namely $c\{0003\}$, $m\{30\bar{3}0\}$, $J\{0.3.\bar{3}.12\}$, $L\{03\bar{3}9\}$, must be written in the multiple form to meet the rhombohedral requirement.

TABLE 1. UNGEMACHITE: FORMS AND ANGLES
Hexagonal— R ; Rhombohedral— $\bar{3}$

$$a:c=1:2.2966; \quad \alpha=62^\circ 51\frac{1}{2}'; \quad p_0:r_0=2.6519:1; \quad \lambda=108^\circ 15\frac{1}{2}'$$

Forms	Bravais	Miller	No. of Faces	Measured				Calculated	
				Range		Mean		ϕ	ρ
				ϕ	ρ	ϕ	ρ		
Basal pinakoid:									
<i>c</i>	0003	111	14	—	—	—	0°00'	—	0°00'
Hexagonal prism of the first order:									
<i>m</i>	3030	211	4	—	—	30°00'	90 00	30°00'	90 00
Hexagonal prism of the second order:									
<i>a</i>	1120	101	16	—	—	0 00	90 00	0 00	90 00
Positive rhombohedrons of the first order:									
<i>b</i>	1.0.1.13	544	1	—	—	30 00	11 30	30 00	11 32
<i>d</i>	2.0.2.23	977	1	—	—	30 00	12 56	30 00	12 59½
<i>e</i>	1.0.1.10	433	2	—	—	30 00	14 49	30 00	14 51
<i>f</i>	2.0.2.17	755	1	—	—	30 00	17 10	30 00	17 19½
<i>g</i>	1017	322	15	—	17'	30 00	20 45½	30 00	20 45
<i>k</i>	4.0.4.19	955	1	—	—	30 00	29 25	30 00	29 10½
<i>h</i>	1014	211	19	—	15	30 00	33 31	30 00	33 32½
<i>i</i>	2025	311	7	—	15	30 00	46 45	30 00	46 41½
<i>j</i> °	4047	511	3	—	—	30 00	56 40	30 00	56 34½
<i>r</i>	1011	100	28	—	24	30 00	69 21	30 00	69 20½
Negative rhombohedrons of the first order:									
<i>D</i>	0.1.1.11	443	5	—	23	−30 00	13 33½	−30 00	13 33½
<i>E</i>	0118	332	3	—	37	−30 00	18 26	−30 00	18 20½
<i>F</i>	0.2.2.13	553	2	—	—	−30 00	22 05	−30 00	22 11½
<i>G</i>	0.4.4.23	995	1	—	—	−30 00	24 50	−30 00	24 45½
<i>H</i>	0115	221	7	—	10	−30 00	27 55	−30 00	27 56½
<i>I</i>	0.4.4.17	773	1	—	—	−30 00	32 00	−30 00	31 58
<i>J</i>	0.3.3.12	552	1	—	—	−30 00	33 17	−30 00	33 32½
<i>K</i>	0227	331	8	—	10	−30 00	37 05	−30 00	37 09
<i>L</i>	0339	441	4	—	26	−30 00	41 41	−30 00	41 28½
<i>M</i>	0112	110	27	—	51	−30 00	52 53	−30 00	52 58½
<i>N</i>	0221	111	16	—	23	−30 00	79 19	−30 00	79 19½
Right rhombohedron of the second order:									
<i>p</i>	1123	210	21	—	17	0 00	56 50½	0 00	56 51
Left rhombohedron of the second order:									
<i>P</i>	2113	201	13	—	20	60 00	56 51	60 00	56 51

TABLE 1.—(Continued)

Forms	Bravais	Miller	No. of Faces	Measured				Calculated	
				Range		Mean		ϕ	ρ
				ϕ	ρ	ϕ	ρ		
Positive right rhombohedrons of the third order:									
<i>g</i>	2134	310	3	6	7	10 51	60 20	10 53½	60 18½
<i>s</i>	3145	410	1	—	—	16 06	62 31	16 06	62 23½
<i>t</i>	2131	201	2	18	6	10 45	81 57	10 53½	81 53
Negative right rhombohedron of the third order:									
<i>U</i>	1322	121	8	18	11	—49 05	74 06	—49 06½	74 05½
Negative left rhombohedrons of the third order:									
<i>V</i>	1232	211	6	15	12	—10 51	74 08	—10 53½	74 05½
<i>W</i>	1235	320	2	24	7	—10 40	54 35	—10 53½	54 31½

The unit rhombohedron $\{10\bar{1}1\}\{100\}$ was chosen to correspond to the unit cell of the structural lattice. The geometrical form of the lattice being known, criteria were apparent in the gnomonic projection which would have led to the same choice had they been clearly recognized. The most direct is afforded by the characteristic interrupted Bravais symbol series in the main radial zone $[c\ m]$, of which more will be said on another occasion. Two further indications can be obtained by analogy with the usual distribution of forms on a simple lattice. In such cases the axial zones, $[a\ b]$, $[b\ c]$, $[c\ a]$, are normally the strongest zones while the radial zone-bundles through the poles of the planes of the unit pyramidal form are the strongest that can be drawn. In the rhombohedral lattice, which is a simple lattice, the axial zones are those joining the upper three poles of the primitive rhombohedron and forming an equilateral triangle whose center is at the center of the projection; the poles of the unit form $\{11\bar{1}\}$ lie at the apices of the inverted circumscribed equilateral triangle whose sides are twice as long as those of the primitive triangle. Inspection of the projection (Fig. 5) shows that with these criteria we would choose $r\{10\bar{1}1\}\{100\}$ as the unit rhombohedron since the zones through the three poles of r , each with ten face-points, are the strongest of their kind. Independently we would select $N\{02\bar{2}1\}\{11\bar{1}\}$ as the unit "pyramidal" form since its poles are the nodes of the strongest zone-bundles through first order rhombohedral face-points.

With this choice of primitive rhombohedron the mean measurements on the common forms lead to the reciprocal (polar) elements:

$$p_0:r_0 = 2.6519:1; \lambda = 108^\circ 15\frac{1}{2}',$$

giving the direct (linear) elements:

$$a:c = 1:2.2966; \alpha = 62^\circ 51\frac{1}{2}'.$$

Table 1 gives a summary of the measurements on eleven crystals of ungemachite together with the two-circle angles calculated from the above elements. The generally excellent agreement between the mean measured and calculated angles show that the crystals approach geometrical perfection with unusual closeness.

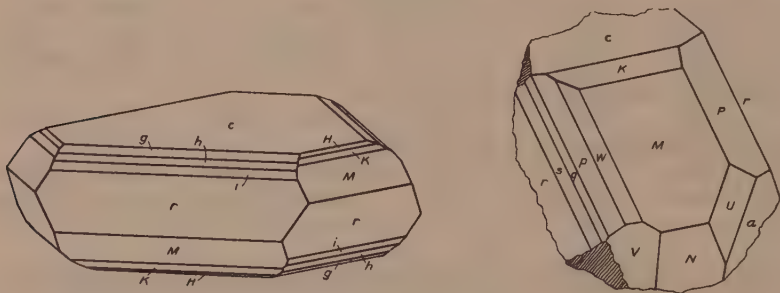


FIG. 6 (left). Ungemachite. Crystal of usual habit.

FIG. 7 (right). Ungemachite. Crystal fragment showing unusual development of third order rhombohedral planes.

The crystals of ungemachite are very uniform in habit (Fig. 6); they are usually bounded by a large base and complex series of positive and negative rhombohedrons of the first order, accompanied by a short prism of the second order and the right and left rhombohedrons of the second order. Strong development of rhombohedrons of the third order was observed only on one crystal fragment (Fig. 7), which is based on a sketch and measurements by Professor Palache. The most important forms are $r c M$; next in importance are $p h N a g P K U H i V$; the rare forms are $D m L j E q e F t W$; the accessory forms, each seen but once, are $b d f k G I J s$.

Elements and Content of the Structural Cell. The following x-ray photographs with copper radiation were made on a small crystal of ungemachite: basal Laue; rotation about the three-fold axis; zero-layer and second-layer Weissenberg about the three-fold axis. The Laue photograph clearly showed the symmetry of the three-fold axis and the absence of vertical symmetry planes, in keeping with the rhombohedral class. The rotation photograph showed eleven serviceable layer lines giving a good value for the principal lattice period, c_0 . The zero-layer Weissenberg photograph showed the trigonal symmetry with very weak reflections from the first, third, fifth and seventh orders and strong reflections from the second, fourth and sixth orders of $(11\bar{2}0)$ which were used for determining a_0 .

Reciprocal lattice projections of the Weissenberg photographs showed only the points of the rhombohedral lattice ($h+i+l=3n$, hexagonal notation). As there are only two space groups in the rhombohedral class (hexagonal tetartohedry of the second kind—Schoenflies), the hexagonal space group $C_{3i}^1-C\bar{3}$, and the rhombohedral space group $C_{3i}^2-R\bar{3}$, the space group of ungemachite is definitely determined as $C_{3i}^2-R\bar{3}$. It is interesting to note that the morphology alone, which clearly reveals the rhombohedral lattice, also gives the space group in the present favorable case.

The elements of the hexagonal (triple) cell are:

$$a_0 = 10.84 \pm 0.02 \text{ \AA}, c_0 = 24.82 \pm 0.05 \text{ \AA}; a_0:c_0 = 1:2.290.$$

The ratio of the cell edges agrees well with the morphological ratio, $a:c = 1:2.2966$. The calculated elements of the rhombohedral (simple) cell are:

$$a_{rh} = 10.37 \text{ \AA}; \alpha = 62^\circ 59\frac{1}{2}'$$

in which the rhombohedral axial angle compares closely with the morphologically determined angle, $\alpha = 62^\circ 51\frac{1}{2}'$. The volume of the rhombohedral cell is:

$$V_{rh} = 842 \text{ cubic \AA}$$

The density of ungemachite, measured by flotation, is:

$$d = 2.287 \pm 0.003$$

The molecular weight of the rhombohedral cell is therefore:

$$M_{rh} = 1167$$

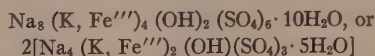
TABLE 2. UNGEMACHITE: ANALYSIS AND CELL CONTENT

$$M_{rh} = 1167$$

	1	2	3	4	
Na ₂ O	21.61	22.15	21.6	Na	8.34
K ₂ O	11.35	11.63	12.3	K	2.88
Fe ₂ O ₃	7.69	7.88	7.0	Fe'''	1.15
SO ₃	40.23	41.23	41.8	S	6.01
H ₂ O	16.69	17.11	17.3	H	22.18
N ₂ O ₅	trace	—	—	O	36.28
Insoluble	2.07	—	—		—
	99.64	100.00	100.0		

A sample of ungemachite was prepared by Dr. Berman and analyzed by Mr. Gonyer, with the result given in Table 2, in which column 1 gives the reported percentage weights, column 2 the percentage weights reduced to 100 per cent after deducting the insoluble portion, column 4

the number of each kind of atom in the unit cell, as obtained from the determined molecular weight of the cell. Column 3 gives the percentage weights corresponding to the deduced cell formula, $K:Fe'''$ being taken as 3:1. The close approach to whole numbers in the last column is satisfactory. The unit cell of ungemachite therefore contains:⁵



in which $K:Fe'''$ is about 3:1. To be sure that ferric oxide is not derived from the very slight amount of impurity contained in the sample, a minute crystal fragment, which appeared water-clear and quite free from foreign substance under the microscope, was tested for ferric iron; the result was strongly positive.

The composition of ungemachite does not compare with that of any sulphate described in Groth (1908) or Mellor (1922); in a system of mineralogy it will therefore appear with the basic hydrous alkali sulphates without close relatives.

Physical and Optical Properties. Ungemachite has a perfect and easy basal cleavage. The hardness is $2\frac{1}{2}$. The crystals are brittle, breaking with an irregular fracture showing a glassy luster. The mineral is uniaxial, optically negative with strong double refraction: $\omega(Na) = 1.502 \pm 0.002$, $\epsilon(Na) = 1.449 \pm 0.002$. Except for slight turgid inclusions the crystals dissolve completely in weak HCl.

CLINO-UNGEMACHITE

Among the crystals selected for measurement were a few that were outwardly indistinguishable from ungemachite but gave gnomonic projections of a distorted trigonal character. The distortion was such that the three poles corresponding to the upper faces of the rhombohedron $r\{100\}$ of ungemachite formed an isosceles triangle, two of the poles having a polar distance $2\frac{1}{2}^\circ$ greater than the third, which lay sensibly at the same angle as in ungemachite. Furthermore a feeble zone was found whose axis was inclined a degree or so to the vertical. Several crystals of this sort were found, each showing precisely the same amount of departure from the rhombohedral form, an amount many times greater than the greatest variation in the measured angles of ungemachite.

⁵ In a summary of the properties of ungemachite and clino-ungemachite (1936) the cell formula is in error, (SO_3) being written instead of (SO_4) ; this error was corrected in a second notice (1937). The erroneous formula has unfortunately been repeated by Spencer (1937 A, 1937 B).

As pointed out by Dr. Berman, the structural equivalence of K and Fe''' implied by this formula is improbable; this would be avoided by writing the cell content as:

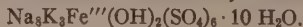


TABLE 3. CLINO-UNGEMACHITE: FORMS AND ANGLES
Monoclinic— F ; Prismatic— $2/m$

$$a:b:c=1.6327:1:1.7308; \quad \beta=110^{\circ}40'$$

$$r_2:p_2:q_2=0.6175:0.6546:1; \quad \mu=69^{\circ}20'$$

Forms	No. of Faces	Measured		Calculated	
		ϕ_2	ρ_2	ϕ_2	ρ_2
<i>c</i> 001	4	69°20'	90°00'	69°20'	90°00'
<i>b</i> 010	1	— —	0 00	— —	0 00
<i>a</i> 100	6	0 00	90 00	0 00	90 00
<i>m</i> 110	2	0 01	33 10	0 00	33 12½
<i>d</i> 012	2	69 18½	50 52	69 20	51 00
<i>e</i> 011	2	69 17½	31 43½	69 20	31 42
<i>f</i> 101	2	33 21	90 00	33 30½	90 00
<i>g</i> 102	3	100 43	90 00	100 43	90 00
<i>h</i> 101	2	127 05	90 00	127 05	90 00
<i>k</i> 113	1	52 56	65 20	52 57	65 16½
<i>p</i> 111	5	33 28	46 18½	33 30½	46 18
<i>l</i> 331	1	14 23	36 50	14 49	36 57½
<i>n</i> 113	1	90 08	60 00	90 01½	60 01
<i>o</i> 111	4	127 05	35 55½	127 05	35 54½
<i>q</i> 331	1	161 49	31 31	161 41½	31 30½
<i>r</i> 315	2	106 48	71 36½	106 50	71 40
<i>s</i> 131	2	33 23½	19 14½	33 30½	19 13½
<i>t</i> 10.1.10	1	127 02	81 58	127 05	82 08½
<i>u</i> 313	1	127 06	65 22	127 05	65 17
<i>v</i> 131	1	127 02	13 26	127 05	13 34½
<i>w</i> 533	1	23 45	55 06	23 49	55 03
<i>x</i> 211	1	20 08	58 17	20 43½	58 31
<i>y</i> 311	1	161 37	61 30	161 41½	61 28
<i>z</i> 811	1	6 10	79 42	6 03	79 40

Professor Palache examined such a projection and saw that it represented a tilted monoclinic crystal which could be placed in a possible crystallographic position by setting the slightly inclined vertical axis upright. Two monoclinic crystals were then measured and projected with reference to the two-fold symmetry axis, which corresponds to the edge (100)/(111)=[011] of the pseudo-rhombohedron. It was then found

that a preferable setting would be obtained by swinging the pseudo-rhombohedral forward about the two-fold symmetry axis until the upper front face stood vertically, the pseudo-rhombohedral faces (100), (010), (001), (111), receiving, in simplest terms, the monoclinic symbols (100), ($\bar{1}11$), ($\bar{1}\bar{1}1$), (001), respectively. In this setting the elements, forms and angles of clino-ungemachite are those given in Table 3 and represented graphically in Fig. 8.

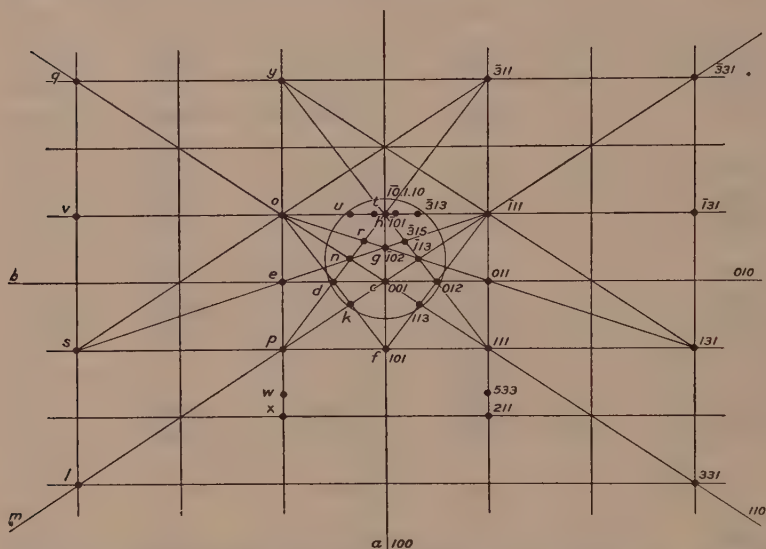


FIG. 8. Clino-ungemachite. Gnomonic projection of the observed forms showing the characteristic absence of planes due to all-face-centering of the lattice.

Later study of the geometrical relations of ungemachite and clino-ungemachite showed that the chosen setting of clino-ungemachite results in an all-face-centered monoclinic lattice cell, as shown in Fig. 9. The reversible transformations, with reference to the pseudo-rhombohedral lattice as the simple crystal lattice, are:

Pseudo-rhombohedral to monoclinic: $2\bar{1}1/01\bar{1}/011$

Monoclinic to pseudo-rhombohedral: $\frac{1}{2}0\frac{1}{2}/0\frac{1}{2}\frac{1}{2}/0\frac{1}{2}\frac{1}{2}$

It is clear that the equivalence of planes given in the previous paragraph is in reality:

Pseudo-rhombohedral		Monoclinic
(100)	=	(200)
(010)	=	($\bar{1}11$)
(001)	=	($\bar{1}\bar{1}1$)
(111)	=	(002)

and that the monoclinic symbols should conform to the F -lattice centering law: h, k, l , all odd or all even. Table 3, in which the indices are retained in their simplest form, shows that the symbols of the majority of the general planes do conform to the above condition. Figure 8 likewise shows the systematic omission of points which do not meet the F -condition, more clearly in the outer field of the projection where the first layer reciprocal lattice pattern is unobscured by gnomonic points which properly belong to higher reciprocal lattice levels.

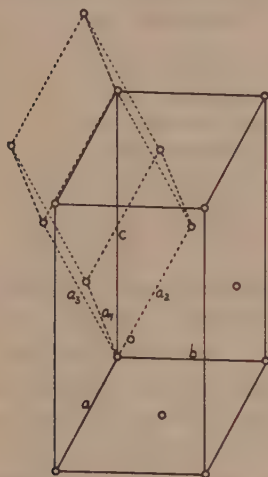


FIG. 9 (left). Clino-ungemachite. Morphological lattice showing the pseudo-rhombohedral simple cell (axes $a_1 a_2 a_3$) in relation to the all-face-centered monoclinic cell (axes $a b c$).

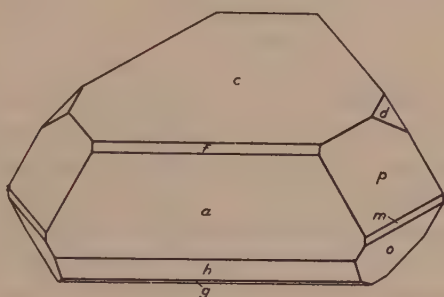


FIG. 10 (right). Clino-ungemachite. Typical pseudo-rhombohedral crystal.

Since an all-face-centered monoclinic lattice can always be referred to a one-face-centered monoclinic cell, the effect of such a transformation was considered in the present case. It was found, however, that a change either to the A -cell or to the C -cell resulted in a setting of the crystal which concealed the striking resemblance to a crystal of ungemachite. The F -lattice cell is therefore retained.

As in the case of ungemachite the agreement between calculated and measured angles is very good, even in the case of the single observations. Of the twenty-four forms on clino-ungemachite, seventeen correspond to observed forms on ungemachite; the common forms on clino-ungemachite are also important forms on ungemachite, and thus a typical monoclinic crystal (Fig. 10) looks much like an average rhombohedral crystal tilted forward in the described manner.

In Table 4 are given the symbols and angles for some of the principal corresponding faces on the two species. The differences in the corresponding angles range from half a minute between *cr* on ungemachite and *ca* on clino-ungemachite to $4^{\circ}46\frac{1}{2}'$ between *cM* and *ch*, clearly proving the crystallographic individuality of the monoclinic species.

TABLE 4. UNGEMACHITE AND CLINO-UNGEMACHITE:
CORRESPONDING SYMBOLS AND ANGLES

Ungemachite	Clino-ungemachite
<i>c</i> (111): <i>h</i> (211) = $33^{\circ}32\frac{1}{2}'$	<i>c</i> (001): <i>f</i> (101) = $35^{\circ}49\frac{1}{2}'$
: <i>r</i> (100) = $69\ 20\frac{1}{2}$: <i>a</i> (100) = $69\ 20$
: <i>H</i> (122) = $27\ 56\frac{1}{2}$: <i>g</i> ($\bar{1}02$) = $31\ 23$
: <i>M</i> (011) = $52\ 58\frac{1}{2}$: <i>h</i> ($\bar{1}01$) = $57\ 45$
: <i>M</i> (110) = $52\ 58\frac{1}{2}$: <i>p</i> (111) = $54\ 06\frac{1}{2}$
: <i>N</i> (11 $\bar{1}$) = $79\ 19\frac{1}{2}$: <i>m</i> (110) = $78\ 51\frac{1}{2}$
: <i>r</i> (010) = $69\ 20\frac{1}{2}$: <i>o</i> ($\bar{1}11$) = $71\ 45\frac{1}{2}$

In all, only six minute crystals of clino-ungemachite were detected; and of these only one remained after some unsuccessful attempts were made to obtain some of their physical and optical properties. There was, therefore, no possibility of analyzing the substance, and the description necessarily remains incomplete. At the same time it seems desirable to place these observations on record so that the mineral may be recognized and more fully described if it should be found again in larger amounts.

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THE ANTHOPHYLLITE AND CUMMINGTONITE-GRUNERITE SERIES

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Madison, Wisconsin.*

Since the publication of a second study of the amphiboles,¹ about seven years ago, much new information of importance has become available. One of the most important contributions is to be credited to Sundius,² who analyzed and measured the optical properties of carefully chosen samples of cummingtonite and grunerite and derived the relationships between optical properties and composition much more accurately than had previously been possible. In a later study³ Sundius discussed the chemical relations of stable forms in the anthophyllite, cummingtonite and actinolite series of amphiboles, and concluded that anthophyllites and cummingtonites are not dimorphous, since he considered that anthophyllites contain less than 40 per cent of ferrous iron, while cummingtonites (and grunerites) contain 40 per cent or more. Bowen and Schairer⁴ reported the results of a study of a remarkably pure iron grunerite and showed graphically the relations between optical properties and composition in natural amphiboles of the cummingtonite-grunerite series as well as in artificial fluor-amphiboles of similar composition.

Warren,⁵ Sundius,⁶ Tilley and Flett,⁷ Simpson,⁸ Tilley,⁹ Rao,¹⁰ Eskola,¹¹ and Wayland¹² have described certain samples of cummingtonite

¹ Winchell, A. N., Further studies in the amphibole group: *Am. Mineral.*, vol. **16**, pp. 250-266, 1931.

² Sundius, N., The optical properties of manganese-poor grunerites and cummingtonites compared with those of manganiferous members: *Am. Jour. Sci.*, vol. **21**, pp. 330-344, 1931.

³ Sundius, N., Ueber die Mischungslucken zwischen Anthophyllit-Gedrit, Cummingtonit-Grunerit und Tremolit-Aktinolith: *Zeits. Krist.*, B(M.P.M.), vol. **43**, pp. 422-440, 1933.

⁴ Bowen, N. L., and Schairer, J. F., Grunerite from Rockport, Massachusetts, and a series of synthetic fluoramphiboles: *Am. Mineral.*, vol. **20**, pp. 543-551, 1935.

⁵ Warren, H. V., *Mineral. Mag.*, vol. **22**, p. 477, 1931.

⁶ Sundius, N., *Geol. För Förh. Stockholm*, vol. **56**, p. 98, 1934.

⁷ Tilley, C. E., and Flett, J. S., *Min. Abst.*, vol. **4**, p. 402, 1931.

⁸ Simpson, E. S., *Jour. Roy. Soc. W. Australia*, vol. **18**, p. 61, 1931-32.

⁹ Tilley, C. E., *Mineral. Mag.*, vol. **24**, p. 181, 1935.

¹⁰ Rao, B. Rama, *Mysore Geol. Dept.*, **15**, 1934.

¹¹ Eskola, P., *C. R. Geol. Soc. Finlande*, vol. **9**, p. 475, 1936.

¹² Wayland, R. C., *Am. Mineral.*, vol. **21**, p. 607, 1936.

and grunerite, and Henderson,¹³ Simpson,¹⁴ Orlov,¹⁵ Ignatiev,¹⁶ Sundius,¹⁷ and Eskola¹⁸ have described samples of anthophyllite and gedrite. Also, under the name "amosite," Peacock¹⁹ described in 1928 samples of asbestos from South Africa, which he classified as orthorhombic amphiboles.

At the time of the previous study, data were insufficient to permit a graph showing in the anthophyllite series the variation in the optic angle, and the curve for N_m was merely a suggestion. New data made it possible for Sundius to draw curves for N_m and the optic angle from 0 to about 40 per cent of the iron end-member. At present the data of Peacock permit the extension of the curves for N_o and N_p , and the specific gravity to the pure iron end-member by extrapolation only beyond 90 per cent, as shown in Fig. 1.

In the cummingtonite-grunerite series Simpson⁸ described a "ferruginous kupfferite" with only 23 numerical per cent of the $\text{H}_2\text{Fe}_7\text{Si}_8\text{O}_{24}$ end-member and an extinction angle of 15° . He has very generously sent to the writer a portion of the material that served for the analysis. A study of this material by Reginald G. Comer, using the double variation method of Emmons, has shown that it consists of two minerals, one having parallel extinction and the other having a maximum extinction angle of 18° in the vertical zone. The second mineral is decidedly uncommon and forms at most about ten per cent of the whole. Therefore, the analysis represents approximately the composition of an orthorhombic mineral, and the composition of the monoclinic mineral is unknown. This orthorhombic mineral has two cleavages (or partings) parallel with the elongation, but they are parallel and normal to the optic plane and must therefore be pinacoidal in position. After careful search, the poorly developed prismatic cleavages were found also, at an angle of about 125° with each other and bisected by the optic plane. This orthorhombic amphibole is clearly an anthophyllite; it has the following optical properties as measured on two crystal fragments:

$Z \wedge c = 0^\circ$, $(+)$ $2V = 79\frac{1}{2}^\circ$, $N_o = 1.6462$, $N_m = 1.6345$, $N_p = 1.6268$ (calc.); $N_o - N_p = 0.0194$ (D line).

$Z \wedge c = 0^\circ$, $(+)$ $2V = 81^\circ$, $N_o = 1.6462$, $N_m = 1.6345$, $N_p = 1.6265$ (calc.); $N_o - N_p = 0.0197$ (D line).

The dispersion ($F-C$) is 0.0083 for N_o (and also for N_m) on the first crystal, and 0.0086 on the second crystal.

¹³ Henderson, E. P., *Am. Mineral.*, vol. 16, p. 563, 1931.

¹⁴ Simpson, E. S., *Jour. Roy. Soc. W. Australia*, vol. 17, p. 137, 1931.

¹⁵ Orlov, A., *Cent. Min.*, p. 269, 1932.

¹⁶ Ignatiev, N. A., *Min. Abst.*, vol. 6, p. 419, 1937.

¹⁷ Sundius, N., *Zeits. Krist.*, B(M.P.M.) vol. 43, p. 422, 1933.

¹⁸ Eskola, P., *C. R. Geol. Soc. Finlande*, vol. 9, p. 475, 1936.

¹⁹ Peacock, M. A., *Am. Mineral.*, vol. 13, p. 241, 1928.

These properties are almost exactly those of an anthophyllite with 23 numerical per cent of $H_2Fe_7Si_8O_{24}$, as shown in Fig. 1.

The monoclinic mineral likewise has two cleavages (or partings) parallel with the elongation, but they are parallel and normal to the optic plane and must therefore be pinacoidal (100 and 010), just as in the orthorhombic mineral. Only by careful search were the prismatic

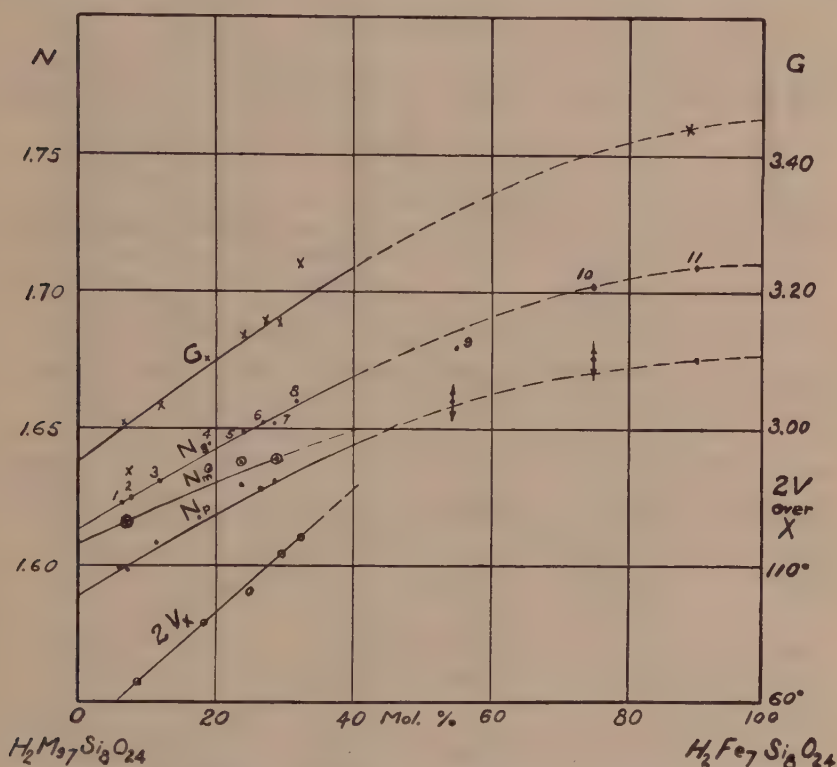


FIG. 1. Variations in composition and properties in the anthophyllite series.

cleavages found, because they are so inferior to the pinacoidal. They are so poorly developed that their position is somewhat uncertain, but they seem to be symmetrically placed about the optic plane at about the amphibole angle. The optic properties of this mineral are:

$Z \wedge c = 18^\circ$, $(+)2V = 78^\circ$, $N_o = 1.6508$, $N_m = 1.6358$, $N_p = 1.6261$ (calc.); $N_o - N_p = 0.0247$ (D line).

Unless actinolite can be optically positive (a condition thus far unknown), these properties indicate that the monoclinic amphibole belongs

at 1a in Fig. 2, and therefore is in the cummingtonite series with about 30 numerical per cent of the $\text{H}_2\text{Fe}_7\text{Si}_8\text{O}_{24}$ end-member.

The extinction angle (15°) reported by Simpson seemed impossible until the discovery by Bowen and Schairer⁴ that in fluor-amphiboles the

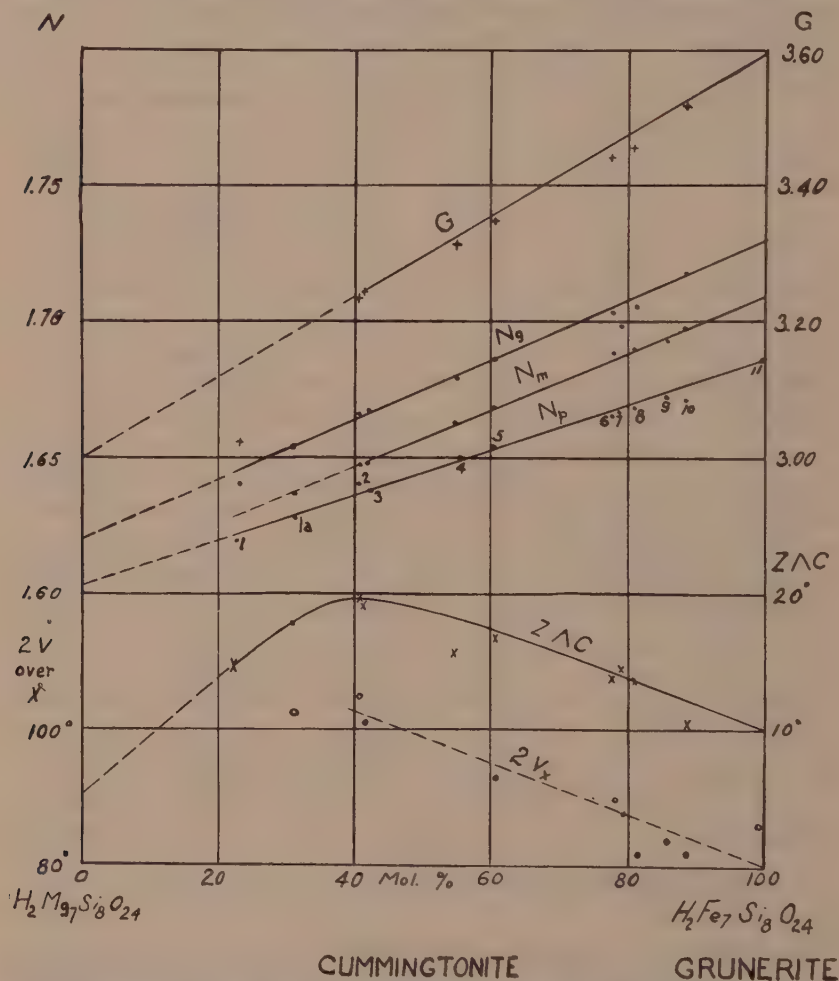


FIG. 2. Variations in composition and properties in the cummingtonite-grunerite series.

extinction angle passes through a maximum at about forty numerical per cent $\text{F}_2\text{Fe}_7\text{Si}_8\text{O}_{22}$. If the extinction angle curve remains about parallel in the two series of amphiboles, natural (hydroxy-) amphibole of this series, containing no iron, would have an extinction angle of about 5° , and this is the value that is derived by extrapolation in Fig. 2.

Perhaps the most important conclusion to be derived from these studies is that the anthophyllite and cummingtonite series actually illustrate a case of isodimorphism, since the cummingtonite series extends beyond 60 numerical per cent of $\text{H}_2\text{Mg}_3\text{Si}_8\text{O}_{24}$ and the anthophyllite series extends at least to about 90 numerical per cent of $\text{H}_2\text{Fe}_7\text{Si}_8\text{O}_{24}$.

This study has been aided by a grant from the Wisconsin Alumni Research Foundation.

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THE OCCURRENCE, CRYSTAL HABIT AND COMPOSITION OF THE URANINITE FROM THE RUGGLES MINE, NEAR GRAFTON CENTER, NEW HAMPSHIRE

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During 1935 the Whitehall Company of New York opened a new pit in the old Ruggles pegmatite, at the top of the mountain where the pegmatite is 300 feet wide. Formerly this mineral deposit had been opened at lower elevations by a series of pits, open cuts, shafts and drifts, in search for mica. The location is $1\frac{1}{2}$ miles N. 40° W. from Grafton Center ($71^{\circ}59'33.6''$ W. Long., $43^{\circ}35'22.3''$ N. Lat., Cardigan quadrangle, New Hampshire). The locality is easily reached by a road; the distance from the village is three miles.

During the operations for feldspar, which occurs as an extensive and unusually pure deposit of this material, the operators occasionally en-

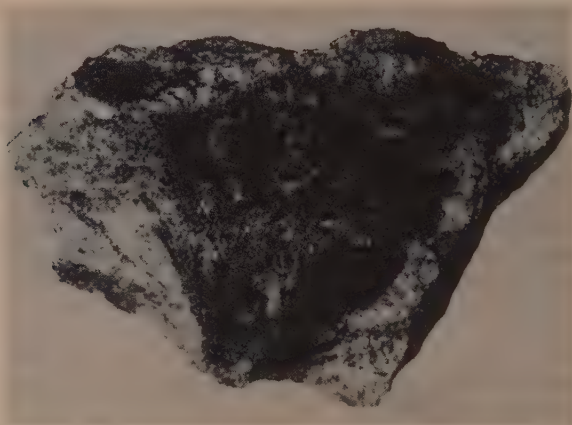


FIG. 1. Uraninite occurring as a three-dimensional dendritic intergrowth in albite-rich perthite. The intergrowth has a fan-shaped directional development. $X=0.25$.

countered remarkable segregations of uraninite and other minerals. The uraninite occurs in all stages of alteration from pitchy black crystals to an impalpable yellowish-white powder, together with an intermediate suite of uranium-bearing minerals which are distinct alteration products. An appreciable quantity of this material was rescued from the dumps and it is hoped that eventually a sufficient amount of the several uranium-bearing minerals may be recovered pure enough to make a detailed study of their physical and chemical properties.

In an occasional specimen of the segregated material the uraninite occurs in small well developed crystals, except for the points where they

were originally attached. The faces on a few of these were lustrous enough to give reflections on the two-circle goniometer. It is the purpose of this paper to describe at this time the occurrence, crystallography and chemical constitution of the uraninite.

MODES OF OCCURRENCE OF THE URANINITE

The usual and most common occurrence of the uraninite consists of a three-dimensional dendritic intergrowth in an albite-rich perthite, and also in massive albite into which the perthite grades. The cleavelandite variety of albite is absent in the part of the pegmatite where the uraninite is found. The intergrowths are often distinctly directional in charac-

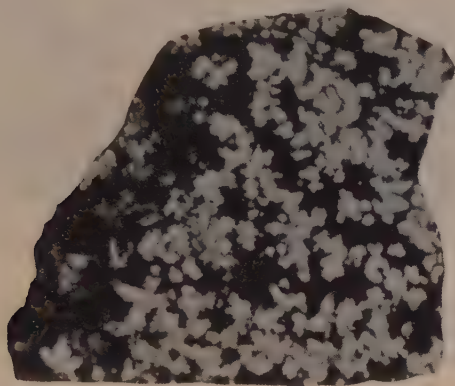


FIG. 2. Radiogram of dendritic uraninite in feldspar.

ter, developing at times into a somewhat fan-shaped structure, Fig. 1. In these directional intergrowths the uraninite crystals are usually joined as is shown by the radiogram, Fig. 2. The crystals occasionally appear to occur as individuals in some cross-sections, in other parallel sections, however, they are probably very often connected. Another frequent occurrence of the uraninite in the perthite consists of spheroidal or ellipsoidal knots consisting of a three-dimensional intergrowth with the feldspar. Both of these occurrences are usually completely surrounded by solid perthite in all directions. The perthite rarely contains individual crystals. One of these, the largest crystal found, consists of a very rough and distorted rhombic dodecahedron modified by a trapezohedron, Fig. 3A.

The pronounced directional uraninite-feldspar intergrowth frequently occurs as a part of a directional asymmetric banding, starting from the normal perthite through an albite-rich perthite with intergrown uraninite, to an apatite-albite intergrowth which meets the normal perthite

again at the opposite side of the banding. The apatite-albite intergrowth is likewise a directional development and often contains uraninite crystals within the albite.

Occasionally the dendritic uraninite in the perthite becomes dense enough to have the appearance of being massive.

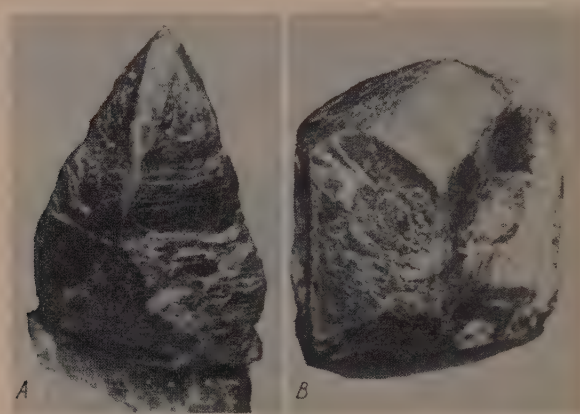


FIG. 3 (A) A distorted rhombic dodecahedron of uraninite with a trapezohedron. The crystal was completely surrounded by perthite. $X=2.55$. (B) Rhombic Dodecahedron of altered uraninite from an apatite-albite intergrowth. $X=4.5$.

Another three-dimensional uraninite-feldspar intergrowth occurs in which there is no apparent directional development of the uraninite, which, in this instance, appears as distinct and sharply outlined skeleton crystals intimately intergrown with massive to medium or fine-grained albite. Some parts of the skeleton crystals are very long in comparison to their thickness, as is shown by the radiogram, Fig. 4.

Two other relationships of the uraninite are noteworthy. One is that of an intricate intergrowth with muscovite in which the uraninite may occur as irregularly developed crystals which sometimes have skeletonized structures. A minor amount of feldspar and quartz may be associated with the muscovite and uraninite. The muscovite in these occurrences is decidedly darkened. The other occurrence of the uraninite is with the quartz of the massive smoky quartz dikes which cut the feldspar, or with the large blebs of smoky quartz in the massive perthite. In these instances the quartz is decidedly darker than elsewhere. The smoky color is undoubtedly due to the effect of being irradiated¹ through

¹ Holden, E. F., The cause of color in smoky quartz and amethyst: *Am. Mineral.*, vol. 10, pp. 221-222, 1925.

the radioactive properties of uranium and its alteration products.

The occurrence of uraninite within the apatite is not common and in such instances the apatite usually shows a radial fracturing around the uraninite.

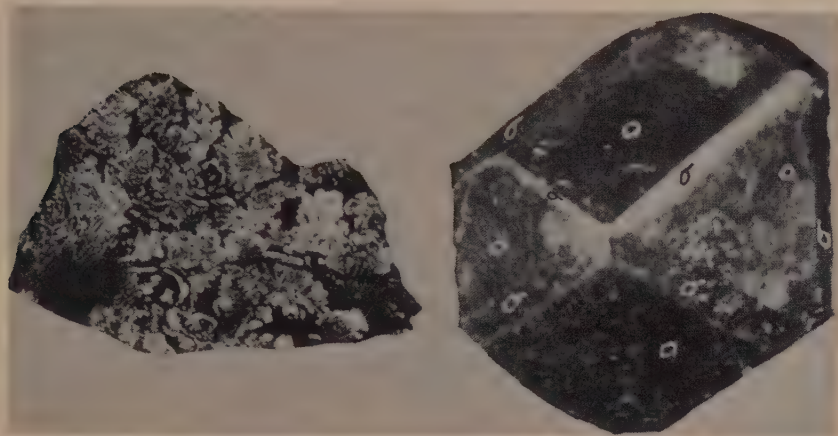


FIG. 4. Radiogram of skeleton uraninite crystals in massive to medium-grained albite.

FIG. 5. A typical crystal of uraninite from the medium to fine-grained albite in the apatite-albite intergrowth. a = cube, o = octahedron, σ = trapezohedron (533). $X=16$.

CRYSTAL HABIT OF THE URANINITE

In the apatite-albite intergrowth the uraninite occasionally occurs as well developed euhedral crystals except for the points of attachment before they were completely surrounded by the albite. The crystals range in size from 1 to 10 millimeters and are often fresh with bright, smooth faces. They are bounded by the cube and octahedron, both about equally developed; in addition a third form, a trapezohedron is common. A typical crystal is shown in Fig. 5. The measured angles between the cube and the trapezohedron is $40^{\circ}13'$, an average of 19 measurements, while the angle between the octahedron and the trapezohedron is $14^{\circ}28'$, an average of 22 measurements. The computed angles between the cube and trapezohedron, (533) is $40^{\circ}19'$, and between the octahedron and trapezohedron (533) is $14^{\circ}25'$. The good agreement between the measured and computed angles establishes a new form, trapezohedron (533), for uraninite.

In another apatite-albite intergrowth a distorted rhombic dodecahedron was found; it is shown in Fig. 3B. This form (110) is extremely rare.

Parsons² has described a trapezohedron on the uraninite from Cardiff Township, Haliburton County, Ontario. His best average determinations on rough crystals with a contact goniometer were 21° between the cube and trapezohedron, and $35^\circ 16'$ between the octahedron and the trapezohedron. The same angles for the form (411) are $19^\circ 28'$ and $35^\circ 16'$, respectively.

The Grafton Center uraninite adds a new form to those already known. This makes a total of five forms which are the cube, octahedron, rhombic dodecahedron and trapezohedrons (411) and (533).

CHEMICAL COMPOSITION OF THE URANINITE

The freshest crystals obtained from the albite of the apatite-albite intergrowth were broken down in stages and the small amount of intergrown impurities, together with pieces which appeared to be altered in any way, were removed under the binocular microscope. From this material 82 milligrams were sent to Friedrich Hecht in Vienna for a microchemical analysis. His determinations are as follows:

Insoluble residue.....	0.54%
SiO ₂	0.80
PbO.....	3.63
(Pb).....	(3.37)
Fe ₂ O ₃	0.66
Al ₂ O ₃	0.44
MnO.....	0.09
Rare earths.....	0.21
ThO ₂	0.43
(Th).....	(0.38)
U ₃ O ₈	90.06
(U).....	(76.38)
CaO.....	0.81
MgO.....	0.17
P ₂ O ₅	0.16
S.....	0.04
H ₂ O (-100).....	0.73
Loss on ignition (100°-1000°C.)....	1.74
Total.....	100.51

The analysis shows the mineral to be a very pure uraninite having an unusually low amount of thorium and rare earths. In this respect it is more like pitchblende which occurs in the metalliferous veins than like uraninite occurring in pegmatites. In the latter occurrences the thorium and rare earths frequently range from 5 to 10 per cent.

² Parsons, A. L., Crystal habit of uraninite from Cardiff Township, Ontario: *Univ. Toronto Studies, Geological Series*, no. 32, pp. 17-18, 1932.

AGE OF THE URANINITE

If we assume that the 0.04 per cent of sulphur present is combined with ordinary lead, the lead-uranium ratio becomes 0.041. By using the simplified logarithmic formula³ one obtains an age of 304 millions of years for the uraninite. According to Holmes⁴ this is late Devonian unless a substantial part of the lead is ordinary lead. As the Ruggles pegmatite is located in an area of highly metamorphosed rocks, any data regarding the age of the intrusives is of interest especially to those working with the stratigraphy of the region. A tabulation of the previously reported lead-uranium and lead-thorium ratios for New England together with the ages determined from these by using the logarithmic formula, is given below:

AGE DETERMINATIONS FOR SOME NEW ENGLAND LOCALITIES BY MEANS OF
LEAD-URANIUM AND LEAD-THORIUM RATIOS

Locality	Minerals Analyzed	No. of Ratios Avg'd	Avg. of Ratios	Age in Millions of years	
				Logarithmic formula	Corrected for Ac-U*
Bedford, N. Y.	Cyrtolite (<i>Munck</i>) ^a	7 for Pb 9 for U	0.051	376	361
Branchville, Conn. (Fillow Quarry)	Uraninite (<i>Hillebrand</i> 3 Anal.) ^b	5	0.052	383	368
Do	(<i>Comstock</i> , 1 Anal.) ^c				
Do	(<i>Boltwood</i> , 1 Anal.) ^d				
Rock Landing, East Haddam, Conn.	Uraninite (<i>Hecht</i>) ^e	1	0.040	297	269
Glastonbury, Conn. (Spinelli Quarry)	Samarskite (<i>Wells</i>) ^f	1	0.039	291	280
Portland, Conn. (Hale Quarry)	Monazite (<i>Fenner</i>) ^g	1	0.037	276	266
Do	Uraninite (<i>Hillebrand</i>) ^h	5	0.039	291	280
Strickland Quarry	Uraninite (<i>Hecht</i>) ⁱ	7	0.041	304	293
Fitchburg, Mass.	Uraninite (<i>Hecht</i>) ^j	1	0.050	370	356

(Table concluded on next page)

* Lane, A. C., *Report of the committee on the measurement of geological time*, National Research Council, Washington, D.C., April 27th, 1935, p. 2.

⁴ Holmes, Arthur, *The Age of The Earth*, p. 176, Thomas Nelson and Sons Ltd., London and New York, 1937.

Blueberry Mt., Mass.	Allanite (<i>Ellsworth</i>) ^k	1	0.043	310	298
Topsham, Maine Do	Monazite (<i>Kroupa</i>) ^l	1	0.079	574(?)	554(?)
	Samarskite (<i>Gonyer</i>) ^m	1	0.031	232	224
Grafton Center, N. H. (Ruggles pegmatite, pit at top of Mt.)	Uraninite (<i>Hecht</i>)	1	0.041	304	293

* Holmes, Arthur, *The Age of the Earth*, p. 153, Thomas Nelson & Sons Ltd., London and New York, 1937.

^a Munech, O. B. The analysis of cyrtolite for lead and uranium: *Am. Jour. Sci.*, vol. 21, pp. 350-357, 1931.

^b Hillebrand, W. F., On the occurrence of nitrogen in uraninite and on the composition of uraninite in general: *Am. Jour. Sci.*, (3), vol. 40, pp. 384-394, 1890; also in *U.S.G.S., Bull.* 78, p. 64, 1891.

^c Comstock, W. J., On the chemical composition of the uraninite from Branchville, Conn.: *Am. Jour. Sci.*, (3), vol. 19, pp. 220-222, 1880.

^d Boltwood, B. B., On the ultimate disintegration products of the radio-active elements: *Am. Jour. Sci.*, (4), vol. 23, pp. 78-88, 1907.

^e Hecht, Friedrich, and Kroupa, E., *Report of the Committee on the Measurement of Geologic Time*, p. 61, May 2, 1936, National Research Council, Washington, D.C.

^f Wells, R. C., *Report of the Committee on the Measurement of Geologic Time*, p. 76, April, 1935, National Research Council, Washington, D.C.

^g Fenner, Clarence N., The age of a monazite crystal from Portland, Conn.: *Am. Jour. Sci.*, vol. 23, pp. 327-333, 1932.

^h Hillebrand, W. F., On the occurrence of nitrogen in uraninite and on the composition of uraninite in general: *U.S.G.S., Bull.* 78, p. 62, 1891.

ⁱ Foye, W. G., and Lane, A. C., Correlation by radioactive minerals in the metamorphic rocks of southern New England: *Am. Jour. Sci.*, vol. 28, p. 133, 1934.

^j Lane, A. C., Age of the Fitchburg granite: *Science*, vol. 78, p. 435, 1933.

^k Holmes, Arthur, *Bull.* 80, p. 343, National Research Council, Washington, D.C., 1931, cites analysis by H. V. Ellsworth in the *Report of the Committee on the Measurement of Geological Time by Atomic Disintegration*, April 28, 1928. Issued privately.

^l Kroupa, E., *Report of the Committee on the Measurement of Geologic Time*, p. 58, May 1, 1937, National Research Council, Washington, D.C.

^m Gonyer, F. L., *Report of the Committee on the Measurement of Geologic Time*, p. 60, May 1, 1937, National Research Council, Washington, D.C.

SUMMARY

The Grafton Center uraninite is unique in occurring chiefly as a three dimensional dendritic intergrowth with perthitic feldspar, and also as skeleton crystals in the same material but more especially in albite intimately associated with the perthite. Individual crystals are associated chiefly with a granular albite intergrown with apatite and also sparingly within the feldspar. A new crystal form, trapezohedron (533) occurs on the uraninite associated with the albite and apatite. The com-

position of the uraninite is chiefly the oxides of uranium, the rare earths and thorium are very low. The lead-uranium ratio is 0.041 which gives it an age of 304 millions of years which is late Devonian.

ACKNOWLEDGMENTS

The writer wishes to express his appreciation and thanks to Dr. A. C. Lane for reading the manuscript and for suggestions which have been incorporated in the table.

ORIENTATION OF BIXBYITE ON TOPAZ

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Berkeley, California.*

Through the courtesy of Mr. John Melhase the writer was permitted to examine and measure a fine specimen of topaz and bixbyite from the Thomas Range, Utah. The specimen is of particular interest because it differs from material hitherto described from this locality.¹ As may be seen from the photograph, Fig. 1, many small bixbyite crystals are attached to one topaz crystal in a horizontal girdle. Closer examination reveals that the bixbyite crystals, which have a cubic habit, (100) modified by (211), as described by Montgomery, have a common orientation, or rather, as appeared after measurement, two complementary orientations.

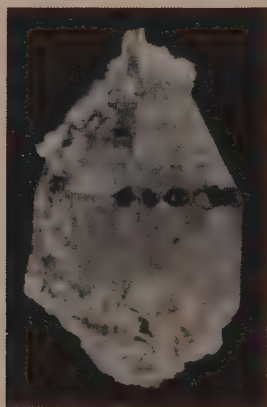


FIG. 1. Crystals of bixbyite oriented on topaz. $\times 3$.

Using a two-circle goniometer, measurements were made on both, topaz and bixbyite at a single setting. The pale amber colored topaz crystal shows only the forms (110), (120), (111) and (021), referred to the axial ratio, 0.5285:1:0.9539, which is in accord with the unit-cell dimensions.

The bixbyite crystals are all attached to the prism faces of topaz, mostly to (110), and reach a maximum dimension of about one millimeter. Since they show mosaic structure and some slight variation of orientation, they could not be measured together. By using an auxiliary magnifying lens attached to the goniometer telescope and a slit dia-

¹ Montgomery, Arthur, A recent find of bixbyite and associated minerals in the Thomas Range, Utah: *Am. Mineral.*, vol. 19, pp. 82-87, 1934.

TABLE 1. TWO-CIRCLE GONIOMETER MEASUREMENTS ON BIXBYITE
REFERRED TO AXES OF TOPAZ

Measured			Calculated			
			turned 32°50'		turned 33°42'	
Bixbyite no. 1	ϕ	ρ	ϕ	ρ	ϕ	ρ
turned clockwise						
100	93° 8	88°36	90°—	90°—	90°—	90°—
001	1 12	31 56	0 —	32 50	0 —	33 42
1 $\bar{1}$ 2	78 16	24 55	76 19	24 51	74 32	25 4
2 $\bar{1}$ 1	99 12	55 50	98 28	55 38	97 54	55 31
1 $\bar{2}$ 1	139 46	38 45	138 42	38 13	137 58	37 34
211	56 56	83 6	55 21	83 1	55 16	83 30
Bixbyite no. 2						
turned counter-clockwise						
100	89° 7	90° 1	90°—	90°—	90°—	90°—
010	0 41	57 43	0 —	57 10	0 —	56 18
121	40 13	38 42	41 18	38 13	42 2	37 34
211	80 2	56 16	81 32	55 38	82 6	55 31
112	101 6	25 0	103 41	24 51	105 28	25 4
2 $\bar{1}$ 1	124 14	84 26	124 39	83 1	124 44	83 30

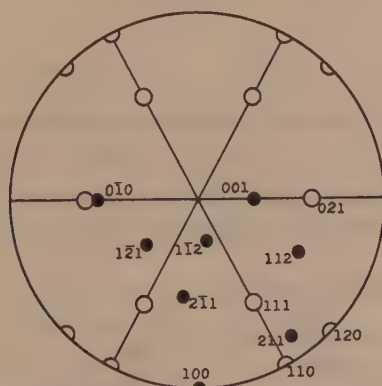


FIG. 2. Stereographic projection of bixbyite crystal oriented on topaz. Black circles indicate poles of bixbyite faces.

phragm to isolate the reflections from a single minute crystal, it was possible to obtain fairly complete sets of observations on six bixbyite crystals, not in contact with each other nor obscured by the adhering matrix of small colorless quartz crystals. All angles for bixbyite were referred to the axial system of topaz. These measurements showed that the [100] axes of the bixbyite crystals nearly coincide with the [100]

axis of the topaz crystal and that a cube face of bixbyite, say (001), is inclined about $32^{\circ}53'$ (average of 6 determinations) to the right or the left from the basal face of the topaz.

Figs. 2 and 3 show the relation of the bixbyite to topaz in stereographic projection and in a front elevation.

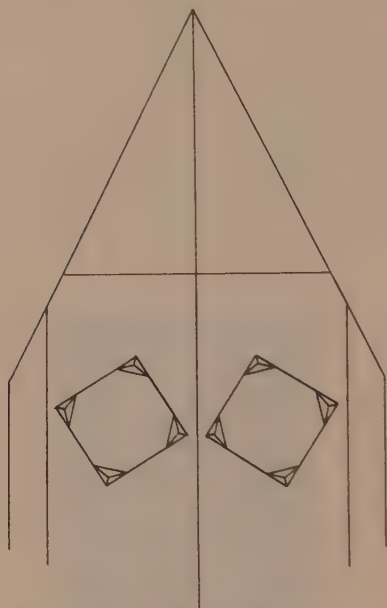


FIG. 3. Front elevation of bixbyite crystals in complementary orientations on topaz.

In Table 1 are given the two-circle measurements on two bixbyite crystals in the positions shown in the front elevation of Fig. 3, together with the angles calculated for bixbyite with $[100]$ coincident with $[100]$

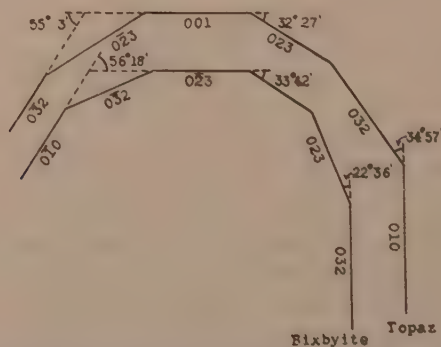


FIG. 4. Correspondence of certain interfacial angles of topaz and bixbyite.

of topaz, and 001 turned clockwise or counterclockwise on the [100] axis either $32^{\circ}50'$ or $33^{\circ}42'$ from the horizontal position. These turning angles were selected from structural considerations explained below.

From Fig. 4 it can be seen that the observed orientation very nearly brings the 010, $0\bar{2}3$, 001 and 032 faces of bixbyite into parallelism with the 032, 001, 023 and 010 faces of topaz.

Formerly this might have been a satisfying explanation, but Royer² and others have shown that regular intergrowths and overgrowths of two minerals are dependent upon the correspondence of spacing and arrangement in certain planes of the structures. Fortunately the structures of bixbyite³ and of topaz⁴ have been fully determined and it is possible to find a structural explanation of this regular overgrowth.

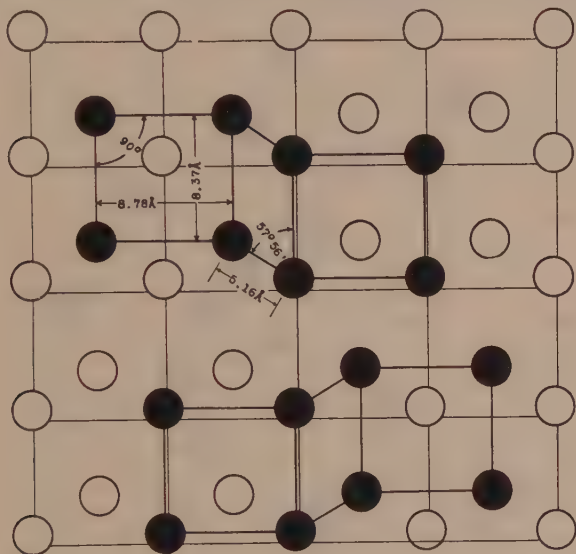


FIG. 5. Arrangement of oxygen atoms in the general position parallel to the 100 face of topaz.

Figure 5 shows the arrangement of oxygen atoms in the general position in a plane parallel to (100) in 16 unit cells of topaz. Figure 6 shows the arrangement of oxygen atoms in a plane parallel to (100) in 16 unit cells of bixbyite. The arrangements of certain groups of eight atoms is

² Royer, M. L., Recherches expérimentales sur l'épitaixie ou orientation mutuelle de cristaux d'espèces différentes: *Bull. Soc. Fran. Min.*, vol. 51, pp. 7-151, 1928.

³ Pauling, Linus, and Shappell, M.D., The crystal structure of bixbyite and the C-modification of the sesquioxides: *Zeits. Krist.*, vol. 5, pp. 128-142, 1930.

⁴ Alston, N. A., and West, J., The structure of topaz, $[\text{Al}(\text{F}, \text{OH})_2]\text{SiO}_4$: *Zeits. Krist.*, vol. 69, pp. 149-167, 1928.

emphasized by blacking in the drawings to show the extent of the correspondence between the two structures. Inspection of the drawings will show that all oxygen atoms in these or similar planes belong to similar groups.

A comparison of the dimensions entered on the diagrams will show that turning the bixbyite lattice some 33° or 34° , either clockwise or counterclockwise, will bring one or another of the marked groups into near-coincidence with a similar group in topaz. The discrepancy in the fit is well within the limits of tolerance found by Royer for the effect of orientation.

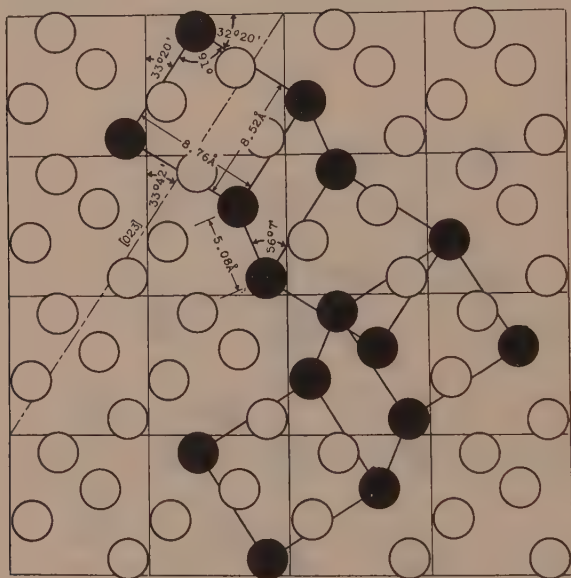


FIG. 6. Arrangement of oxygen atoms in the 100 plane of bixbyite.

If the bixbyite is turned just $33^\circ 42'$ its $[023]$ axis will coincide with the c axis of topaz. The identity distance along the $[023]$ or $[03\bar{2}]$ axis of bixbyite is 33.77\AA , at which distance the configuration of atoms in question is repeated. This matches closely $4c$, 33.51\AA , and $4b$, 35.13\AA , of topaz.

This is equivalent to matching the area of the 100 plane across 13 cells ($3^2 + 2^2 = 13$) of bixbyite, 1140\AA^2 , over the area of the 100 plane across 16 cells of topaz, 1179\AA^2 . The discrepancy is only 3.3%. In these areas 8 out of 52 atoms of oxygen in a 100 plane of bixbyite would coincide with 8 out of 32 oxygen atoms in a 100 plane of topaz.

If, on the other hand, the bixbyite is turned just $32^{\circ}50'$, the nearest approach to coincidence of a single group of 8 atoms with a similar group in topaz is obtained. Inspection of Table 1 and consideration of the average angle of turning observed suggest that this is the governing factor in the orientation.

NOTES AND NEWS

ON THE IDENTITY OF AUSTINITE AND BRICKERITE

WOLFGANG BRENDLER, *Hamburg, Germany.*

In 1932 (*N. Jahrb. Mineral.*, Abt. A, Beil.-Band 66, p. 44, 1932) Friedrich Ahlfeld called attention to an arsenate of zinc and calcium from Lomitos, Bolivia. A complete description was to be given later by its discoverer, Dr. J. Barrande-Hesse of La Paz, Bolivia, who named it "brickerite."

At the fifteenth meeting of The Mineralogical Society of America at Rochester, N. Y., December 27, 1934, Lloyd W. Staples of Stanford University, presented a paper on a new arsenate mineral from Gold Hill, Utah, and gave a complete description of this new mineral, named by him "austinite" in honor of Professor Austin F. Rogers (*Am. Mineral.*, vol. 20, pp. 112-119, 1935).

The chemical analysis (by R. B. Ellestad) of the new mineral that occurs with adamite in minute but distinct, well developed, colorless crystals of bladed or acicular habit, found in the oxidized zone of the Western Utah Copper Company's ore body, lead to the chemical composition CaZn(OH)AsO_4 .

In 1936 (*Zentralbl. Min.*, Abt. A, No. 8, pp. 226-231, 1936) Friedrich Ahlfeld and R. Mosebach described the above mentioned "brickerite" from the Lilli mine near Lomitos, Bolivia. A chemical analysis of the mineral was made by its discoverer, J. Barrande-Hesse. The analyzed material was not pure, but contained 10.46% CaCO_3 , 1.65% SiO_2 and 0.66% Fe_2O_3 . Unfortunately the analyst neglected to determine the water content and therefore arrived at the erroneous chemical formula: $\text{Zn}_4\text{Ca}_3\text{As}_4\text{O}_{17} = 4\text{ZnO} \cdot 3\text{CaO} \cdot 2\text{As}_2\text{O}_5$. Specimens of "brickerite," which I have received from Friedrich Ahlfeld, yield an appreciable amount of water when the powdered mineral is heated in a closed tube, and make it probable, that "austinite" and "brickerite" are identical.

At Lomitos the mineral occurs in banded veins with fibrous structure, resembling fibrous gypsum or aragonite. It is found filling small dikes in a greenish dacite. The fibrous layers alternate with crusts of quartz, chalcedony and limonite. Adamite was not observed on the specimens from Lomitos.

The color of the fresh mineral is light yellowish to colorless. It shows a high lustre and is very brittle. The thickness of the layers varies from a fraction of a millimeter to 6 mm. The color of the powdered mineral is a light yellowish-white. Upon weathering the mineral becomes dull and white; the weathered material contains considerable carbon dioxide and a larger amount of water. The hardness of the fresh mineral is 4-4.5; the specific gravity, 4.13-4.14.

The optical properties of the mineral from Lomitos were given by R. Mosebach and are nearly identical to those found by Lloyd W. Staples on the austinite from Gold Hill, as the following table shows:

Optical character		Lloyd W. Staples positive	R. Mosebach positive
α	=	1.759 ± 0.003	1.752
β	=	1.763 ± 0.003	1.755
γ	=	1.783 ± 0.003	1.779
Maximum double refraction		0.024 ± 0.006	0.027
Optic axial angle $2V\gamma$		ca. 47°	41°

A chemical analysis, made by the writer on carefully selected fresh material, gave the following result:

	I Percent	II Ratio	III	
CaO	21.12	.377	$= 2 \times .188$	+ .001
ZnO	30.51	.375	$= 2 \times .188$	- .001
As ₂ O ₅	42.48	.185	$= 1 \times .188$	- .001
Sb ₂ O ₅	0.53	.002		
P ₂ O ₅	none			
V ₂ O ₅	none			
CO ₂	none			
H ₂ O*	3.47%	.190	$= 1 \times .188$	+ .002
H ₂ O (105°C)	0.12%			
Fe ₂ O ₃	0.69%			
MnO	0.12%			
Insoluble	1.11%			
	<hr/> 99.91%			

* Penfield closed tube method.

The table shows that a small amount of arsenic acid is replaced by antimononic acid. It is strange that a second analysis made by the writer on material from another specimen from the same locality gave no traces of antimony but 42.95% As₂O₅. (This analysis will be published by Friedrich Ahlfeld in a later publication.)

Between the mineral layers, crystals with terminated planes were observed in small cavities. These crystals resemble in habit the spear-shaped marcasite crystals. Unfortunately the faces of the crystals are uneven

and dull so that it is impossible to obtain material for goniometric measurement.

The optical properties and the chemical analysis definitely demonstrate the identity of "austinite" and "brickerite." Priority should be given the name "austinite."

NOTES ON SOME MINERALS FROM SOUTHERN CALIFORNIA

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INTRODUCTION

In this paper have been grouped a number of brief descriptions of new mineral localities in Southern California, and observations on some localities previously described. It is the intention of the writers to publish similar assembled data from time to time, in the hope that in this



INDEX TO LOCALITIES. (1) Crestmore, Riverside County. (2) George's Gap, San Gabriel Mountains, Los Angeles County. (3) Acton, Los Angeles County. (4) Tejon Postoffice, San Joaquin Valley, Kern County. (5) Rincon, San Diego County. (6) San Francisquito Canyon, Los Angeles County. (7) Twenty-nine Palms, San Bernardino County.

way the information may be made more readily available than if issued as separate items. Unless otherwise indicated by initials (M or W) after each locality, the areas described were investigated by both authors.

(1) CRESTMORE, RIVERSIDE COUNTY (M)

Okenite.—Examination of several specimens from the Sky Blue Quarry at Crestmore, Riverside County, showed white, imperfect crystals, composed of silky fibers, and obviously pseudomorphs. These crystals varied from .5 to 1 mm. in diameter, and 2 to 3 mm. in length. A review of the literature on Crestmore, showed that Eakle and Rogers¹ noted these pseudomorphs, and determined them to be okenite after wilkeite. Some of these crystals were well developed, with smooth faces; accordingly they were measured on the goniometer, to determine, if possible, the axial ratio of the original mineral. The prism zone was well developed, and the measured angles checked satisfactorily with the hexagonal prism. No second order prism faces were detected. The terminations in most individuals were well rounded, showing no crystal faces but in several cases, pyramid planes were present. These were measured as accurately as possible, though in only a few instances were signals observed. In all, seven faces were measured, giving ρ angles approximately as follows:

Reading	Angle ρ
1.	40° 30'
2.	40 30
3.	40 30
4.	40 00
5.	40 00
6.	39 30
7.	39 30
<hr/>	
Average	40°04'

This average checks very closely with Eakle and Rogers' reading of 40°09' for a single pyramid face of wilkeite, and further confirms their findings of an axial ratio for this mineral that is close to apatite. If this is assumed as the unit pyramid, $c = 0.730 \pm$.

The pseudomorphs were found in a granular aggregate of okenite, idocrase, and a dark brown material which in part formed the matrix between larger, subidiomorphic grains of idocrase. In addition to the pseudomorphs, okenite also occurs in massive fibrous form, interstitial to idocrase individuals. The dark brown material occurs often in rounded grains, occasionally showing a crystal face, but in no case with recognizable forms. It has a hardness of 2-3, and the microscope shows it to be quite completely altered to a mixture of limonite and a finely fibrous or granular doubly refracting mineral, whose index of refraction is approximately 1.52. It is too fine-grained, and too intimately mixed with

¹ Wilkeite, a new mineral of the apatite group, and okenite, its alteration product, from southern California: *Am. Jour. Sci.* (4), vol. 37, p. 262, 1914.

limonite to permit its identification, but judging by the general shape of the grains, the parent mineral might have been diopside.

Grossularite Garnet.—Blue-green grossularite garnet was also found in the contact zone of the Sky Blue quarry. This occurs as deep to pale, blue-green grains and irregular patches, up to 4–5 mm. across, with occasional dodecahedral crystals up to 1 mm. in diameter. The matrix is massive, colorless grossularite, and the included grains are frequently abundant enough to strongly color the rock. The index of refraction of this garnet is 1.720, and its specific gravity, as determined with Clerici solution is 3.25.

Opal (Hyalite).—In the same locality as the blue-green garnet, and in many cases coating crevices in the garnet rock, is a stalactitic crust. This crust is mainly calcite, but often on it is deposited, in short stalactitic and finger-like forms, a later coating of perfectly clear, colorless opal. This hyalite is on the average 0.5 mm. in diameter and 2–3 mm. long.

(2) ALMANDITE GARNET, FROM GEORGE'S GAP, SAN GABRIEL
MOUNTAINS, LOS ANGELES COUNTY

Almandite Garnet.—Almandite garnet, in excellent, large, euhedral crystals was encountered by the construction crews building a new road in the San Gabriel Mountains, on the Angeles Crest highway, near George's Gap. The almandite crystals are deep brown to reddish-brown in color, and vary from a few millimeters to 3 or 4 cm. across. They occur in a zone of biotite-chlorite schist which is part of the San Gabriel formation as mapped by Miller.² The entire area of the garnet-bearing schist was excavated in building the road, and several hundred large crystals were collected. Many small crystals are present in the surrounding schistose rocks, but none have been found to compare in size with the first. The garnet has a specific gravity of 4.1 as determined with Clerici solution. The dominant form is the dodecahedron, with small modifying trapezohedrons, $n(211)$. When first obtained, specimens were covered with a thick, weathered coating of biotite and chlorite; upon scrubbing, fresh, brilliant, nearly transparent crystals were obtained. The crystal faces are almost invariably pitted due to partial inclusion of chlorite or biotite flakes.

(3) ZEOLITES, FROM ACTON, LOS ANGELES COUNTY (W)

Heulandite.—Excellent, colorless crystals of heulandite, in seams, veins and cavities, were found filling vesicles, joints, and fractures in a blackish gray basalt, located about one-quarter mile west of the village

² William J. Miller, Geology of the western San Gabriel Mountains of California: *Univ. of Calif. at Los Angeles Pub. in Math. and Phys. Sci.*, vol. 1, pp. 1–114, 1934.

of Acton. The lava is a flow occurrence. Specimens are readily obtainable since the basalt has been opened up as a quarry for road material. The crystals vary in size from microscopic forms to 7–8 cm. in length. Veinlets of heulandite appear almost massive, with a cream-colored appearance; upon magnification, these veinlets are seen to be composed of minute crystals, each crystal by itself is nearly colorless. These seams can be traced through the highly fractured rock for some distance, occasionally almost disappearing and again widening into zones an inch or more in width, in which occur the larger crystals. The crystals and vein material check microscopically for heulandite.

Stilbite.—In the same quarry are found two occurrences of stilbite, each with slightly different physical properties. They were both identified as stilbite by optical means. (1) The largest crystals occur in the wider parts of the heulandite seams, and lie upon a base of well-formed heulandite crystals. The stilbite occurs in this relation as milk-white to colorless sheaves and plates, almost rosettes, with exceptionally brilliant pearly luster on all faces. The sheaves are often an inch or more in length; they never completely fill the open spaces in the heulandite seams. (2) A second occurrence of stilbite was noted in the same quarry, where seams and veins, in appearance and physical features like the heulandite seams, transect the basalt. Along the seams are pockets and cavities that are filled with tabular, platy crystals averaging 7–8 cm. in length. These are cream-colored, and often stained with a brown or black coating of iron and manganese oxides. Such cavities are exceedingly irregular, and in many places have been so completely filled that the specimens appear to be brecciated basalt cemented with stilbite. In this second type of occurrence, one never finds any associated heulandite.

Relations between Heulandite and Stilbite.—The heulandite was first deposited in the vein-fractures and joints in the basalt. The perfect crystals indicate freedom during crystallization. Since the stilbite fills the interstices between the heulandite crystals, and rests upon a matrix of them, it must be later, and must have formed well after the heulandite crystallized. The vein occurrence of stilbite was never found in juxtaposition with heulandite, but since optically the two occurrences of stilbite are similar, both occurrences undoubtedly were deposited at the same time.

(4) GIBBSITE, FROM SAN JOAQUIN VALLEY, KERN COUNTY

Gibbsite.—During an investigation of some heretofore little known garnet-bearing rocks on the Tejon Ranch in the Southern San Joaquin Valley, Kern County, some boulders carrying a peculiar clay-like pink mineral were encountered on uplifted alluvial fans on the steep hillslopes

of the Tehachapi Mountains, at the southern edge of the valley. Several boulders were found, though the outcrops from which they were derived could not be located. Examination showed the material to be gibbsite, the hydroxide of aluminum. Index determinations checked, and blowpipe tests gave strong aluminum reaction with cobalt nitrate, and absence of silica. The gibbsite is pale pink to rose-red in color, the deeper shades of pink being most typical. This is a somewhat unusual color for this mineral. The gibbsite occurs in a gray chert breccia, the whole locally stained by limonite. The chert fragments are compactly cemented and partially replaced by the gibbsite, which shows a concentric structure where it fills the interstices; this structure is evident from color variations shown in the intergrowth. Gibbsite is a relatively rare mineral, and this occurrence is the first known to the writers in the state of California, although undoubtedly it is present as an alteration product mixed with the clays of the Riverside County deposits.

(5) GASTALDITE, FROM RINCON, SAN DIEGO COUNTY

Gastaldite.—A bluish fibrous mineral occurring abundantly along the contact of a small pegmatite dike with diorite, was determined by its optical properties to be gastaldite—a variety of glaucophane. A more detailed description of this occurrence is planned to accompany a discussion of the Rincon pegmatites by the writers at a later date.

(6) MARIPOSITE AND CLINOZOISITE FROM SAN FRANCISQUITO CANYON, LOS ANGELES COUNTY

Mariposite.—The rare, green, chromium-bearing mica has been identified from the Sierra Pelona³ schists of the San Francisquito Canyon. The mineral occurs in considerable abundance in the first large canyon entering San Francisquito from the east, just above the upper end of the lake floor that was San Francisquito Reservoir. The mariposite occurs as nests and lenses in talc-sericite and talc-actinolite schists. In some of the larger lenses, the flakes are 10–12 mm. in diameter. In color, the mariposite varies from an emerald-green to a greenish yellow. All of the material gives a chromium bead before the blowpipe. The mariposite is rather widely disseminated in some occurrences, being scattered in small flakes throughout talc schists. The commonest occurrence is in the larger lenses.

A mica closely related to mariposite is reported by Hulin⁴ from the

³ Hershey, Oscar H., Some crystalline rocks of southern California: *Am. Geol.*, vol. 29, pp. 372–390, 1902.

⁴ Geology and ore deposits of the Randsburg Quadrangle, California: *Calif. State Min. Bur., Bull.*, 95, p. 25, 1925.

Randsburg district, where flakes occur in a schist in a similar relation to the San Francisquito Canyon locality. Knopf⁵ described the occurrence of mariposite in the Mother Lode district. However, the San Francisquito canyon occurrence is the first known to the writers in Southern California.

Clinozoisite.—On the eastern edge of what was the San Francisquito reservoir, in the Sierra Pelona schists, near the mariposite locality, occurs a quartz vein in which large crystals of clinozoisite were found intergrown with quartz. The clinozoisite crystals are often three to four inches in length, with bundles of crystals an inch or more in diameter. Much of the clinozoisite outcrop was destroyed in excavating for a road, but good specimen material is still available in the old road embankments.

(7) LAMELLAR QUARTZ FROM TWENTY-NINE PALMS, SAN
BERNARDINO COUNTY, CALIFORNIA (M)

Lamellar quartz.—An outcrop of pure white quartz, about five miles southeast of Twenty-nine Palms, on the road towards White Tank, attracted the writer's attention; investigation showed it to possess some unusual features. Although this particular outcrop could be traced only a few feet in any direction, it was undoubtedly a quartz-rich portion of one of the numerous pegmatite dikes which cut the granite of this area.

The quartz showed in many places a nearly perfect parting, which produced flat surfaces over areas six to eight inches across. The luster on this surface is strongly pearly, and flakes can readily be split off parallel to it. The plane surface is interrupted by occasional rounded and irregular areas of normal glassy quartz, with the usual conchoidal fracture. It is further marked by faint parallel striations, unevenly spaced, sometimes entirely absent, which are seen to be narrow lamellae with slightly different orientation from the main surface. These striations indicate the trace of a second direction of parting, not nearly as perfect at the first, and not so readily visible on a hand specimen. The luster on this second surface is usually dull, as if frosted, but sometimes highly vitreous. In this plane, the edges of the more prominent lamellae of the quartz are plainly visible. The angle between these two directions, measured on a number of specimens, varies between 65° and 75°. Occasionally, a third still poorer direction of parting was observed. Its presence was shown by straight cracks intersecting the best parting surface. This third separation makes an angle of approximately 87° with the first, and 81° with the

⁵ The Mother Lode System of California: *U. S. Geol. Surv., Prof. Paper 157*, pp. 38–39, 1929.

second parting. The luster on the third is strongly vitreous, with the parting surface ordinarily discontinuous over any area, being composed of a number of small, parallel faces.

Several thin sections were made, some cut parallel to the best parting, others normal to it and along the line of the second parting. Those cut parallel to the best, often showed a single large quartz grain (in the area of the section), or sometimes a number of very large grains, in essentially parallel position. Occasional inclusions of differently oriented grains are found, as was observed in the hand specimen. The large grains do not extinguish uniformly, but as a vague mosaic of areas which extinguish almost, but not quite, simultaneously. As nearly as could be observed under these conditions, the c axis of the quartz is inclined at an angle of about 40° to the plane of the parting. Accurate measurement on the universal stage was not possible because of the irregular extinction.

A section cut normal to the best parting, shows long parallel lamellae of varying width, with alternate strips which generally extinguish simultaneously. Neighboring lamellae extinguish only a few degrees apart. Several measurements of this difference gave values ranging from 3° to 5° . These lamellae extinguish uniformly throughout their length, which in some cases is entirely across a section. This confirms the observation on the broadside section, that the lamellae are made up of very large individuals, even though they do not have quite even extinction on the broad face.

The alternate arrangement of these lamellae strongly resembles albite twinning, and the observed structure may be explained as a pseudomorphic replacement of a triclinic feldspar by quartz. The unusual feature of this is the replacement of feldspar lamellae essentially by single individuals, rather than by an aggregate of small grains. The presence of these large individuals is of course responsible for the perfection of the best parting over such large areas.

PROCEEDINGS OF SOCIETIES

MINERALOGICAL SOCIETY OF GREAT BRITAIN AND IRELAND

Meeting held January 27, 1938

DR. L. J. SPENCER, President, in the chair.

DR. C. E. TILLEY exhibited lamprophyllite from the Kola peninsula and molengraaffite from Pilansberg, Transvaal, and gave evidence to prove the identity of the two species. The following papers were read:—

- (1) *The paragenesis of a Malvern hydrobiotite, and the variable content of hydroxyl in micas generally.* By A. BRAMMALL, F. A. BANNISTER, AND J. G. C. LEECH

X-ray and chemical work show that the so-called hydrobiotite is not a mica but a new member of the chlorite family. Its constitution was compared with that of three hydromuscovites, two normal biotites, and a lithionite from Trelavour Down, Cornwall. Dehydration data were given, and experimental work on the hydrolysis of these micas was briefly demonstrated.

- (2) *The Kaalijärvi meteorite from the Estonian craters.* By DR. L. J. SPENCER

The crater-lake of Kaalijärvi on the island of Oesel was first described in 1827, and many suggestions have been made as to its mode of origin. A detailed survey, with excavations and borings, of this and of five other smaller craters was made by Mr. I. Reinvald in 1929 and 1932, and he was convinced of their meteoritic origin, although no trace of meteoritic material could be found. With remarkable persistence he returned to the work of excavation in July, 1937, when he was rewarded by finding thirty small fragments of much rusted meteoritic iron with a total weight of about 100 grams. A polished and etched section of one piece shows much schreibersite in an ataxite groundmass, the latter containing 8.32% of nickel. The meteoritic origin of these craters is therefore now established.

- (3) *Curvature in crystals of vein-quartz.* By DR. A. T. J. DOLLAR

Genetic problems presented by curved and fractured quartz crystals, closely associated with straight and unfractured crystals of the same mineral, are discussed in relation to anomalies of their surface form and internal optical properties. The specimens were derived from a vein in the slates of Lundy Island, Bristol Channel.

- (4) *A petrographic description of lundyite from Lundy Island, Bristol Channel.* By DR. A. T. J. DOLLAR

Lundyite was named by T. C. F. Hall in the Summary of Progress of the Geological Survey of Great Britain for 1914. Two chemical analyses have been made and details of the petrography are brought forward.

NEW YORK MINERALOGICAL CLUB

*American Museum of Natural History, New York City
March 16, 1938*

With Vice-President Dr. Andersen presiding, the meeting was called to order with 55 members and guests present. After some preliminary business and the election of new members, the speaker of the evening, Dr. A. F. Buddington, of Princeton University, addressed the Club upon "Economic Geology of the Karelian-Kola trip of the 17th International Congress." The talk was illustrated with lantern slides.

Dr. Buddington gave an interesting presentation, describing the geology and the unusual deposits of the area. The talk was illustrated by fine maps of the localities and a display of some excellent specimens collected at the localities visited. He complimented the leaders highly upon their careful preparations which made the trip so successful. The personal sidelights that Dr. Buddington was able to give added much to the interest and enjoyment of his audience.

F. H. POUGH, *Secretary*

PHILADELPHIA MINERALOGICAL SOCIETY

Academy of Natural Sciences of Philadelphia, March 3, 1938

A stated meeting of the Philadelphia Mineralogical Society was held on the above date, with the President, Mr. Trudell, in the chair. Forty-three members and twenty-two visitors were present.

Dr. Edward H. Watson, of Bryn Mawr College, spoke on "Some Geological Observations in Mexico." During the spring of 1937, Dr. Watson drove through Mexico, stopping for a few weeks to study the San Carlos Mts. Dr. Watson described huge alkaline laccoliths, some thirty miles wide with a very complex composition. The speaker also drove to Mexico City and described the trip up through the dissected escarpment of the plateau.

Dr. Watson concluded from observations of the Great Valley of Mexico that it is a huge caldera, larger than any yet described. He based his conclusions on structural features, the valley being rimmed by large active and extinct volcanos, smaller vents marking faults within the caldera.

The lecture was interspersed with comments concerning the people and politics of Mexico, and was well illustrated with photographs.

LOUIS MOYD, *Secretary*

Professor Terence T. Quirke of the Department of Geology, University of Illinois, has translated into English two articles by Professor V. M. Goldschmidt. They pertain to the Laws of Rock Metamorphism, with examples from the geology of southern Norway. Students of metamorphic geology who have difficulty in reading German will find these translations very helpful. Mimeographed copies, with paper covers, may be secured from the University of Illinois Supply Store, Inc., Champaign, Illinois, upon payment of a nominal price of sixty-five cents.

